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Report of Activities 2000

Manitoba
Industry, Trade
and Mines

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Geological Survey

REPORT OF ACTIVITIES
2000

2000

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REPORT OF ACTIVITIES 2000

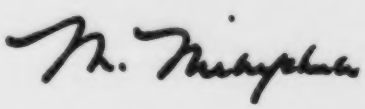
A message from the Minister:

Manitoba is one of the best places in the world to invest in exploration and mining. Our province enjoys many strategic advantages in this sector, not the least of which is its mineral resource endowment. One of our greatest assets is the skill and dedication that staff of the Manitoba Geological Survey bring to unraveling the secrets of Manitoba's mineral potential.

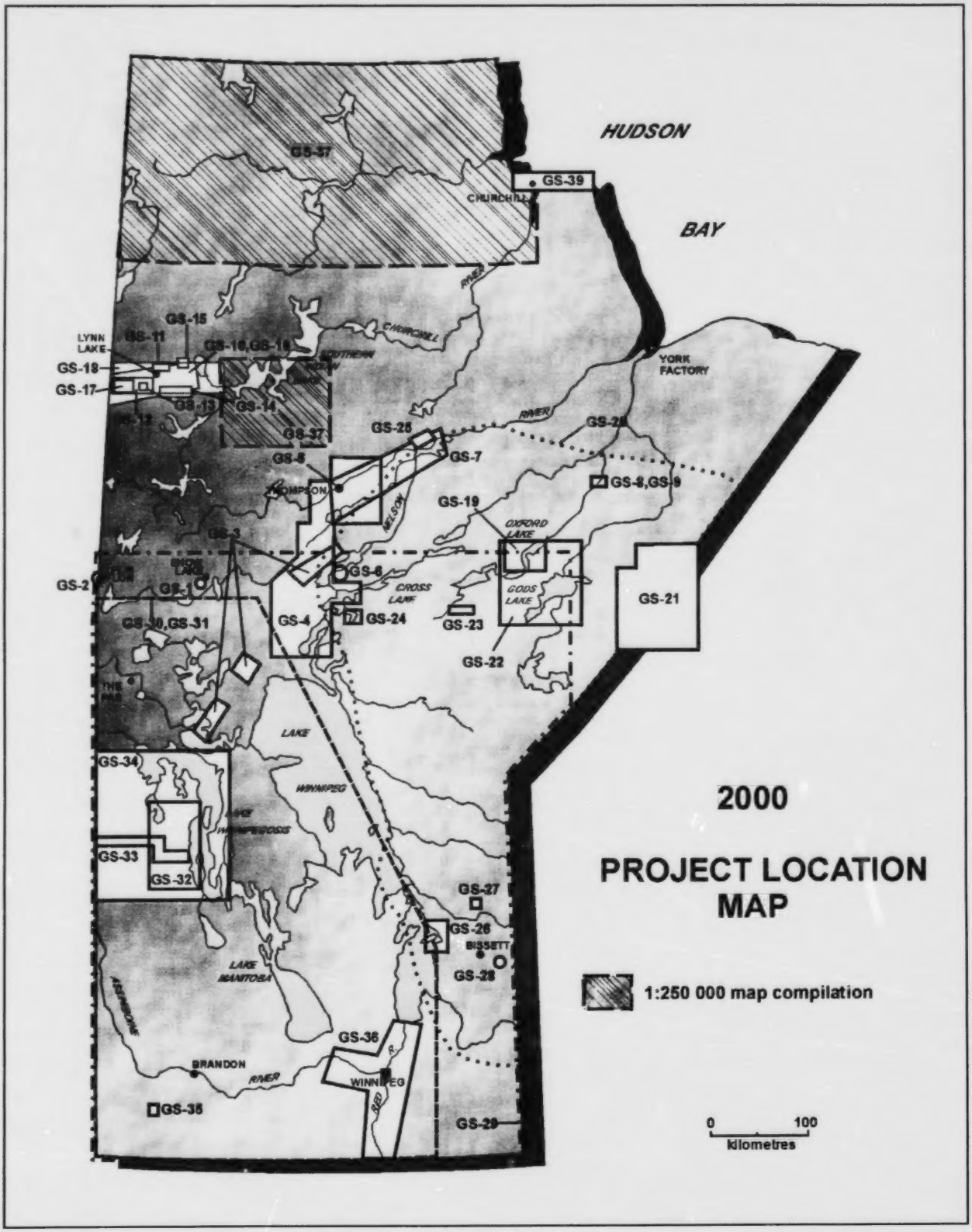
The **Report of Activities** is a comprehensive summary of the geoscience investigations carried out by departmental geological staff during the 2000 field season. It also includes contributions reporting on collaborative projects with a number of Canadian universities, the mining and mineral industry, and various geoscience organizations across Canada. It is our intent to provide the mining and exploration industry with the most current geoscience information possible, to help focus exploration strategies for the coming year.

The geoscience community will note that this year the Geological Services Branch has been renamed the *Manitoba Geological Survey*. This new name more clearly reflects the role of the Branch and is consistent with usage in other provincial geological survey organizations across Canada. The Survey recognizes the importance of the mining industry to the Manitoba economy and is committed, with the rest of the department, to supporting initiatives that ensure the long-term viability of the mining industry and mining communities.

The release of this report at the 2000 Manitoba Mining and Minerals Convention creates an ideal forum to stimulate discussion and produce beneficial consultations between Manitoba Industry, Trade and Mines staff and stakeholders in the Manitoba mining community. The mining and minerals industry has an important role to play in the Manitoba Government's vision for a prosperous and dynamic economy that generates sustainable employment growth for all Manitobans. I hope this report provides you with the solid geoscience base you need to plan your next exploration venture in Manitoba.



Honourable MaryAnn Mihychuk
Minister of Industry, Trade and Mines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by E.C. Syme, Acting Director, Manitoba Geological Survey

Syme, E.C. 2000: Introductory summary; *in* Report of Activities, 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 1-8.

This millennial year marks the 71st year (since 1929) that a Manitoba government agency has conducted geological investigations within its jurisdiction. Annual publications devoted solely to geological investigations in Manitoba began in 1968, making the *Report of Activities 2000* the 33rd annual publication of its kind. *Report of Activities* has grown over time to become a rich source of geoscience information. The steady expansion in scientific content reflects the commitment of Manitoba Geological Survey staff to excellence in their submissions. The increasing involvement of university, federal government and industry partners in our geological investigations has broadened the scope of the activities reported in the pages that follow.

The full text and graphic content of individual reports in the *Report of Activities 2000* can be downloaded, free of charge, from the departmental web site, by pointing your browser to http://www.gov.mb.ca/em/geoscience/geo-index.html and following the links.

GEOSCIENCE PROGRAM

The primary mandate of the Manitoba Geological Survey (MGS) is to provide public sector, accessible, reliable geoscience information — material that is required by industry to attract investment in the increasingly competitive global mineral and petroleum exploration markets. By developing an understanding of Manitoba's geology and geological processes, the MGS also contributes to decisions regarding land-use management. To help meet these objectives, funding for the Manitoba Geological Survey was increased in 2000 to $4.4 million.

Major field-based activities conducted by the MGS include bedrock mapping at a number of scales, as well as more specialized, thematic projects. Compilation maps and regional syntheses (1:50 000 scale or smaller) provide up-to-date regional-scale information for both the major mining districts of Manitoba and the frontier areas of the province. They also provide geological background and mineral-resource assessments in support of land-use planning and sustainable development. Thematic studies (1:20 000 scale or larger) address specific geological and land-use problems, support mineral-deposit studies and compilation-series mapping, and identify new methods of assessing and defining mineral potential for a variety of traditional and nontraditional mineral-deposit types. Specialized geochemical and geophysical projects provide regional data to assist industry in the identification of prospective areas. Maintenance of a mineral-deposit inventory provides a systematic catalogue of metallic-mineral, industrial-mineral and aggregate deposits.

The 2000–2001 geoscience program reflects a balance between providing support to traditional mining camps, stimulating new exploration and development opportunities in frontier areas, and supporting land-use and development priorities in southern Manitoba. The program is reviewed annually by the Mineral Exploration Liaison Committee (MELC), composed of members of the Mining Association of Manitoba, the Manitoba Prospectors and Developers Association, and the Manitoba–Saskatchewan Prospectors and Developers Association, as well as representatives from the University of Manitoba and the Geological Survey of Canada (GSC).

The MGS geoscience program is designed with a regional emphasis, focusing on those areas most in need of new or updated geological information. Programs are generally designed on a cycle that allows the geographic focus of activities to shift from area to area within the province. In 2000–2001, approximately 41% of operating resources are directed toward studies in the northern Superior Province, a reflection of the Department's five-year initiative to stimulate exploration in this underexplored part of the province. Approximately 40% are directed toward projects within traditional mining camps (Flin Flon Belt, Lynn Lake Belt, Thompson Nickel Belt and southeastern Manitoba). The remaining 19% support Quaternary, aggregate, industrial minerals and land-use planning in the south-central part of the province, as well as GIS support for all projects.

Partnership Programs

Geoscience programming in Manitoba is delivered through a combination of provincially funded and collaborative projects. The partnered initiatives include contributions from the federal government, the mineral exploration industry and several Canadian universities (Table 1). These partnerships are anticipated to add an additional $1.0 million to operating field expenditures in 2000/01, effectively matching the amount of operating funds spent on geoscience by the province.

Targeted Geoscience Initiative

A new federal government program aimed at providing additional funding for geoscience in Canada was introduced in the 2000 federal budget. The Targeted Geoscience Initiative (TGI) brings $5 million per year in new federal spending, for three years, to projects across the country. The goal of the TGI is to turn resource potential into new social and economic benefits by increasing the level and effectiveness of private-sector mineral exploration. By improving the quality and quantity of available geoscience information and promoting its rapid dissemination on the Internet, the TGI will help stimulate new investment in the mining sector.

Targeted Geoscience Initiative projects are agreed upon and delivered collaboratively by the GSC and provincial/territorial surveys, based on priorities developed in concert with local technical advisory committees (MELC in Manitoba). Manitoba-based TGI projects approved and in progress to date include:

- **Regional hydrothermal systems as an exploration guide, FlinFlon mining camp, Manitoba and Saskatchewan:** With combined federal and provincial funding of approximately $800 000 over three years, this project will include studies of 1) the mineralogical and stable-isotope characteristics of alteration below and above the Flin Flon–Callinan–777 stratigraphic interval, 2) the nature of regional deformation patterns affecting the Flin Flon VMS horizon, 3) the reconstruction of the nature and distribution of volcaniclastic rocks associated with the Flin Flon VMS horizon, 4) the nature and distribution of hydrothermal alteration within the Bear Lake and Bakers Narrows blocks, 5) regional oxygen-isotope alteration, and 6) the geochemistry of the Flin Flon massive-sulphide deposits. An orientation survey was conducted during a short field season in 2000, but the main work associated with the Flin Flon TGI will commence in 2001.
- **Southeastern Manitoba–Red Lake (Ontario) digital compilation:** With combined federal and provincial funding of approximately $403 000 over three years, this project will provide the necessary resources to collect a variety of geological, geochemical and geophysical data sets, in digital form, for an area in northwestern Ontario and southeastern Manitoba. The aim is to produce a CD-ROM containing all available, integrated geospatial data for the region, to aid all types of mineral exploration in the region from the Red Lake Belt in Ontario to the Rice Lake and Bird River belts in Manitoba. Work on the Southeastern Manitoba–Red Lake TGI will begin in the fall of 2000.

The MGS and the GSC will work together to propose additional projects in priority areas of the province, for years two and three of the Targeted Geoscience Initiative. Given that TGI is a federal program,

Table 1: Collaborative Geoscience Programs in Manitoba, 2000-2001.

Region	Project and partners
Thompson	Thompson Nickel Belt CAMIRO (INCO Ltd., Falconbridge Ltd., HBED, Billiton Metals Canada Inc., Western Mining International Ltd., GSC, U of Manitoba, Laurentian U, U of Alberta, U of Québec at Montréal, U of Saskatchewan, NSERC)
Thompson	Thompson Nickel Belt compilation (INCO Ltd., Falconbridge Ltd., HBED)
Superior BZ	Fox River Belt geology and metallogeny (Falconbridge Ltd.)
Superior BZ, Northern Superior	Western Superior LITHOPROBE (U of Waterloo, U of Alberta, NSERC)
Northern Superior	Western Superior NATMAP (OGS, GSC, universities, LITHOPROBE, NSERC)
Lynn Lake	Lynn Lake gold studies (U of New Brunswick, Laurentian U, U of Manitoba, NSERC)
Snow Lake	Subvolcanic intrusions CAMIRO (now completed; GSC, universities)
Flin Flon Belt	Shield Margin NATMAP (now completed; GSC, SGS, universities, LITHOPROBE)
Flin Flon Belt	Flin Flon Belt geochronology (HBED)
Flin Flon Belt	Flin Flon TGI project (GSC, SGS, industry)
Interlake	Hydrogeology (GSC, WRB, U of Manitoba)
Southeastern Manitoba	Southeastern Manitoba-Red Lake TGI project (GSC, OGS)
Southern Manitoba	Greater Winnipeg NATMAP (GSC, universities)
Southern Manitoba	Paleofloods of the Red River (GSC, IJC, RRFPP)

Abbreviations: CAMIRO, Canadian Mining Industry Research Organization; GSC, Geological Survey of Canada; HBED, Hudson Bay Exploration and Development Ltd.; IJC, International Joint Commission; LITHOPROBE, Canada's National Geoscience Program; NATMAP, National Geoscience Mapping Program; NSERC, Natural Sciences and Engineering Research Council; OGS, Ontario Geological Survey; RRFPP, Red River Flood Protection Program; SGS, Saskatchewan Geological Survey; Superior BZ, Superior Boundary Zone; TGI, Targeted Geoscience Initiative; U, University; WRB, Water Resources Branch.

final approval of projects rests with the GSC. However, independent of the success of future TGI proposals, the MGS will continue to address critical geoscience needs in the province.

Brief summaries of the results obtained during the 2000 field season are provided below, arranged according to the geographic area in which they were conducted.

FLIN FLON–SNOW LAKE

The NATMAP Shield Margin Project, completed in 1998 with the publication of a set of 1:100 000 scale compilation maps and accompanying notes, was the culmination of twenty years of MGS mapping and mineral-deposit investigations in the Flin Flon Belt. The scientific results of the Shield Margin Project have now been published in two volumes of the Canadian Journal of Earth Sciences (January and November 1999). A CD-ROM containing the compilation maps and a variety of other digital data sets (potential field images, geochemistry, geochronology, selected 1:50 000 and 1:20 000 scale maps, field data) was released in May of this year.

With the completion of the Shield Margin Project, the intensity of MGS field investigations in the Flin Flon Belt has decreased. However, the important lessons learned during the Shield Margin Project, in terms of greenstone belt complexity and project methodology, are now being applied in the Western Superior NATMAP project and in the Lynn Lake Belt. Most current projects in the Flin Flon Belt have a specific mineral-deposit orientation (e.g. gold deposits in the Snow Lake area, documentation of VMS deposits from mine plans and new sampling). In the effort to complete the Shield Margin Project, work on a large number of 1:5 000 to 1:20 000 scale maps was put on hold, and we now intend to begin publishing final maps for completed projects. These detailed maps will form the basis of a new 1:50 000 scale compilation series for the belt. As described above, the Targeted Geoscience Initiative projects will bring joint MGS–GSC–Saskatchewan Geological Survey resources to bear on specific base-metal–related topics in the western Flin Flon Belt.

At Snow Lake, an alteration study was undertaken as part of a B.Sc. thesis at the University of Manitoba. This study comprises detailed mapping (1:1 000 scale) and sampling of an outcrop of silicified basalt, and comprehensive mineralogical and geochemical studies. It is designed to fully characterize the nature of the synvolcanic hydrothermal alteration and to provide a comprehensive understanding of the silicification event.

LYNN LAKE–LEAF RAPIDS

A variety of new initiatives began in 2000 in the Lynn Lake region, in response to the closure of the Keystone mine's Farley deposit, formerly the only active mining operation in Lynn Lake. These initiatives include new regional and detailed structural mapping in the Johnson Shear Zone (JSZ) and Agassiz Metallotect, selected multimedia geochemical surveys, and a GIS compilation project that will house not only the newly acquired information but also a wide variety of existing data not readily available at present. An M.Sc. study, currently underway at the University of Manitoba, utilizes existing and new digital data sets in the Lynn Lake area to develop a predictive tool for gold exploration. These initiatives augment MGS office-based compilation initiatives, including production of new 1:50 000 scale colour maps for the belt and updated geochemical data pertaining to the volcanic stratigraphy of the belt.

Gold Metallogeny in the Lynn Lake Belt

Many MGS investigations in the Lynn Lake greenstone belt are intended to improve understanding of processes involved in the formation of gold mineralization in the region. The objective is to provide the exploration community with metallogenic models for the formation of Lynn Lake–type gold deposits. The 2000 field season was the second year of a multiyear, multifaceted gold metallogeny project that represents a significant expansion of MGS efforts in the Lynn Lake Belt. The initial fieldwork, started during the 1999 field season, focused on the deformation history of the Johnson Shear Zone and the relationship to gold mineralization hosted by the shear zone. Fieldwork completed during the 2000 season saw a continuation of the 1999 program and the addition of thesis projects investigating past-producing gold deposits associated with the JSZ and Agassiz Metallotect.

The Johnson Shear Zone is a major east-trending, pre–peak metamorphism shear zone located in the southern Lynn Lake Belt.

Evidence is accumulating for a long history of movement along the shear zone, with the majority of observed shear fabrics representing a major period of reactivation of an older structure. Mapping in 2000 extended the JSZ to the Dunphy Lakes area, west of Gemmell Lake, for a total strike length that is now known to exceed 100 km. Gold mineralization west of Gemmell Lake, associated with sheared and altered felsic volcanic rocks, is characterized by the presence of very fine grained acicular arsenopyrite and local finely disseminated pyrite, zones of syn-shear quartz veins, silicification and weak sericitization. The gold mineralization west of Gemmell Lake, hosted by sheared felsic volcanic rocks, represents a different and previously unrecognized style of mineralization within the JSZ and has excellent gold exploration potential. A second style of mineralization, informally termed 'BT-style' after the Burnt Timber gold deposit, is associated with finely disseminated pyrite in highly silicified, carbonatized, sheared mafic volcanic rocks. The silicification is accompanied by moderate biotitization, reflecting syn-shear potassic metasomatism.

The Agassiz Metallotect is an economically important, broadly stratiform zone of regional extent that hosts significant gold deposits within the northern Lynn Lake Belt. Preliminary mapping and detailed structural analysis in the western portion of the metallotect has identified five generations of structural fabrics formed by discrete deformational events. Stratabound gold mineralization within the metallotect occurs in both metavolcanic and metasedimentary rocks. Associated hydrothermal alteration includes calcic amphibolitization, biotitization, muscovitization, silicification and carbonatization, as well as chloritization and serpentinization of the host high-Mg-Ni-Cr basalt (picrite) and fine-grained sedimentary rocks. Gold mineralization seems to be principally associated with the second-generation carbonate-quartz-sulphide veins crosscutting the sulphidic metasiltstone.

Preliminary structural and geochemical investigations of the Farley Lake gold deposit have determined that the gold mineralization is associated with the emplacement of postdeformational, postplutonism quartz-carbonate-sulphide veins into banded oxide- and silicate-facies iron-formation, resulting in sulphidization of the iron-formation. The high-grade quartz-carbonate-sulphide veins were emplaced along pre-existing, shallowly southwest-dipping joints and subvertical, second-generation faults and shear zones. The thick sequence of iron-formation and interbedded argillaceous sedimentary rocks that hosts the mineralization has experienced significant first- and second-generation fold thickening.

Mineralization-Potential Mapping, Lynn Lake

The construction of a regional database of geological and geophysical data for the Lynn Lake Belt was begun in 1999 as an M.Sc. study at the University of Manitoba. The objective of the project is to produce mineralization-potential maps using a GIS platform and data-fusion technique. Data sets compiled include regional geology, lithogeochemistry, lake sediment and till geochemistry, airborne geophysics, and synthetic aperture radar (SAR) data. A preliminary set of mineralization-potential maps for shear-hosted and VMS-type deposits, produced using theoretical principles, show good correlation with known mineral occurrences throughout the southern portion of the Lynn Lake Belt.

Volcanic Stratigraphy and Geochemistry, Kisseynew North Flank

With the publication of new trace-element data from the Lynn Lake Belt in 1999, additional follow-up work was conducted within the boundary zone between the Lynn Lake Belt proper and the Kisseynew Domain. This work helps to define the extension of Lynn Lake Belt tectonostratigraphic assemblages into the more highly deformed and recrystallized domain to the south. Highly recrystallized, mafic to ultramafic volcanic rocks, previously mapped as undifferentiated amphibolite, can be distinguished geochemically in the same manner as in the better preserved rocks of the Lynn Lake Belt. Volcanic units in the Kisseynew margin have geochemical and metallogenic affinities to volcanic arcs, marginal-basin floor (N-MORB and contaminated MORB) and ocean-island basalt. Despite structural attenuation and alteration involving alkali elements, the content of the least mobile elements in assemblages is remarkably uniform for tens of kilometres along strike and provides a valuable tool for mapping. Major structural breaks separate the arc and marginal-basin assemblages, even where units were attenuated to a thickness of only a few metres. Shear zones that reactivate or cut these breaks may host precious-metal deposits.

New Geochronological Results from the Lynn Lake Belt

Geochronological studies are helping to establish the geological evolution of the Lynn Lake Belt. New U-Pb ages suggest that the bulk of plutonic activity, both mafic and felsic, occurred during the period 1870–1876 Ma. The nickeliferous 'EL' and barren Fraser Lake gabbroic stocks, both of which intrude the Wasekwan Group arc-related metavolcanic rocks in the Lynn Lake area, have yielded U-Pb zircon ages of 1871.3 ± 2.4 Ma and 1870 ± 6.2 Ma, respectively. The Eden Lake granite, intruding gneissic tonalite at the southeast end of the Lynn Lake Belt, yielded an age of 1870 ± 10 Ma that is comparable to the age of the gabbros and to the published ages of the Hughes Lake and Norrie Lake tonalite plutons. The age of a tonalite pluton north of the Fox Lake mine, at the southwest end of the Lynn Lake Belt, is significantly younger at 1831 ± 3.7 Ma. The 1831–1871 Ma period of intrusive activity indicated by this new work is similar to the span of 'successor arc' magmatism determined elsewhere in the Trans-Hudson Orogen.

THOMPSON NICKEL BELT–FOX RIVER BELT

Thompson Nickel Belt

The MGS has turned a significant part of its attention to the Thompson Nickel Belt (TNB), an area that, despite its economic importance to the province, has historically seen relatively little provincial survey work due to limited bedrock exposure and a large amount of hitherto confidential company information. Recognition by all stakeholders that the public geological database for this world-class mineral belt required major updating resulted in two initiatives, both supported by industry and both entering their final stages in 2000–2001:

1) Compilation of the geology of the exposed TNB and its sub-Phanerozoic extension, begun in 1995, has been made possible by the generous co-operation of INCO Ltd., Falconbridge Ltd., and Hudson Bay Mining and Smelting Co. Ltd., allowing MGS access to virtually all private-sector information in the TNB. Preliminary versions of the maps are planned for publication in the spring of 2001.
2) The Thompson Nickel Belt CAMIRO project brings researchers from universities and government together to develop new exploration tools for the TNB in a four-year (1997–2001) study, in part funded by industry (*see* below).

Thompson Nickel Belt bedrock geology compilation

The geological compilation of the TNB progressed this year with the completion of NTS sheets 63F/08, 63F/09, 63G/14 and 63J/03. During the summer, diamond-drill cores from 317 holes from a broad area around Wabowden were re-examined. Thirty per cent of these drillholes contain Ospwagan Group metasedimentary rocks, host rocks to the Thompson-type nickel deposits. This new work is notable because no Ospwagan Group rocks were shown on the 1972 compilation map of this area. Although only one-third of the available core was examined in the summer of 2000, major lithological trends will be compiled onto maps from the newly obtained data. Re-logging of additional drillholes will be undertaken in the summer of 2001 to further refine the compilation map.

In concert with the TNB compilation, petrographic and geochemical studies were conducted to provide detailed mineralogical data for the Ospwagan Group, Kisseynew Domain rock units and reworked Archean gneiss in and adjacent to the Thompson Nickel Belt.

Thompson Nickel Belt CAMIRO Project

The Thompson Nickel Belt CAMIRO Project has a combined industry–NSERC budget of approximately $800 000, and is administered by the Canadian Mining Industry Research Organization (CAMIRO). The MGS co-ordinated and led the project until August of 2000, when personnel changes required that these responsibilities be split between coleaders at the University of Manitoba and Laurentian University. The project currently has five industry sponsors (INCO Ltd., Falconbridge Ltd., Hudson Bay Exploration and Development Ltd., Billiton Metals Canada Inc. and Western Mining International Ltd.) and researchers from the GSC, the University of Manitoba, Laurentian University, the University of Alberta, the University of Québec at Montréal, and the University of Saskatchewan.

Regional mapping in the TNB

At Mystery Lake, a regional syncline has a core of Ospwagan Group metasedimentary and metavolcanic rocks that were intruded by large ultramafic bodies. The Archean basement gneiss on the limbs of the syncline contains high-strain zones that feature kinematic indicators interpreted to be the result of northwest shortening with dominantly sinistral, east-side-up displacement. The Mystery Lake data suggest that northwest compression and vertical extension were dominant, and that megaboudinage may have caused a local reversal in the horizontal shear component. The results do not support a model involving only dextral transpression, but are instead consistent with the structural style of undisrupted, long, sinuous folds that has been identified during preparation of a compilation map of the Thompson Nickel Belt.

The metamorphic transition between the Archean Pikwitonei Granulite Domain and the Paleoproterozoic, amphibolite-grade Thompson Nickel Belt was examined on western Sipiwesk Lake. Fieldwork focused on collecting samples from layers of leucocratic, mesocratic and melanocratic rocks that could be reliably traced across this well defined metamorphic boundary. In addition, a series of samples was collected from a well exposed, layered, mafic-ultramafic complex in the Pikwitonei Domain to study metamorphic reaction textures at granulite grade. Results of these studies will be reported in a B.Sc. thesis currently in progress.

Fox River Belt

Falconbridge Ltd. launched a major Ni-Cu-PGE exploration program in the Fox River Belt (FRB) in 1999. This new exploration work provides the MGS with an opportunity to advance the geological database for the belt, which is the largest continuous segment of the Superior Boundary Zone in Manitoba. The MGS completed a reconnaissance field study of the FRB in 1999, and results obtained from this program provided strong impetus for a multiyear geological investigation of the metallogeny, petrogenesis and tectonic evolution of the belt.

The 2000 field program included reconnaissance geological studies in the western part of the FRB. The Lower Sedimentary formation and the adjacent Archean orthogneiss of the Superior Province host one or more suites of previously unrecognized mafic intrusions. Granitic rocks locally intrude the Lower Sedimentary formation in the southwestern part of the Fox River Belt, and may relate to previously unknown Proterozoic felsic plutonism within the Archean–Proterozoic boundary. In the Fox River Belt, most of the sedimentary rocks are undeformed and consist of repeated cycles, tens of metres thick, of early clastic and later chemical sedimentation. Limited sedimentological studies of the Middle Sedimentary formation and upper parts of the Lower Sedimentary formation suggest a strong similarity to the distal marine sedimentary rocks in the prospective parts of the Ospwagan Group in the Thompson Nickel Belt. The FRB is bordered on the north by paragneiss, with turbiditic greywacke protoliths, that has been folded and metamorphosed at amphibolite facies. This paragneiss was sampled in 2000 to determine its Nd-Sm model age.

The Great Falls area provides the best known exposures of the Fox River sill within the Fox River Belt. Mapping completed in 2000 focused on the Marginal Zone outcrops and enabled completion of a 1:1000 scale geological map of the northern half of the Great Falls outcrop area. Mapping has shown that the lower part of the Fox River sill in this area is largely undeformed and has a chilled contact with older pyritic mudstone and siltstone of the Middle Sedimentary formation. Surface prospecting has delineated disseminated Fe-Cu-Ni sulphide mineralization within the Marginal zone at three different stratigraphic levels. The KO zone is a stratabound zone of sulphide mineralization, enriched in Cu, Ni and platinum-group elements (PGE), that developed immediately above the irregular contact between the host ultramafic rocks of the second cyclic unit and the upper, leucogabbroic part of the first cyclic unit. Recent analytical results confirm the PGE-rich nature of the KO zone, which has a maximum grade of 5.4 g/t combined Pd+Pt+Au, 2.3% Cu and 1.1% Ni.

Ongoing research of the Marginal zone geology and mineralization is being carried out as a Ph.D. thesis at the University of Manitoba. This research will provide new geological, mineralogical and geochemical observations that will allow the development of rigorous genetic models for the emplacement and mineralization of the KO zone. On a regional scale, ongoing detailed investigations of the Fox River sill will continue to provide constraints on its emplacement and crystallization. One hypothesis currently being tested, which is consistent with many recent field and geochemical observations, is that the Fox River sill was emplaced on a subvertical, syn-rift, axial fault system, and is therefore a major dyke.

NORTHERN SUPERIOR PROVINCE

Manitoba's five-year Northern Superior initiative is now in its final year of data collection. The objectives of this initiative are broadly to: 1) identify regional exploration targets through new geochemical surveys and compilations of geophysical data derived from assessment files; and 2) provide the geological framework for mineral exploration through regional mapping and thematic studies. This program is linked to the larger Western Superior NATMAP project that is focused on defining the crustal evolution and tectonic assembly of the Superior Province west of Lake Nipissing.

Multimedia Geochemical Survey, Knee Lake

Operation Superior, begun in 1996, involves helicopter-supported multimedia geochemical surveys targeting greenstone belts in the northern Superior Province. During the 2000 field season, approximately 2000 rock, till, b-horizon, humus and vegetation samples were collected for multi-element geochemical analysis from 350 sites within the northern portion of the Knee Lake greenstone belt. In addition, 334 bulk samples of till and beach sand were collected for identification and analysis of kimberlite indicator minerals. The geochemical and mineralogical analysis of these samples will continue to build a multi-element–multimedia geochemical database to assist mineral-resource assessment in the northern Superior Province of Manitoba. Future directions include a joint data-integration project conducted by the MGS and the GSC, to produce a belt-by-belt multi-element–multimedia product that will highlight and prioritize areas of mineral potential.

Regional Mapping in the Northern Superior Province

Regional (1:20 000 to 1:50 000 scale) bedrock geological mapping is being conducted in selected greenstone belts in the northern Superior Province, including current work in the Knee Lake–Gods Lake, Stull–Kistigan–Edmund lakes, and Max Lake–Aswapiswanan Lake areas. Lithological, geochemical, structural and geochronological data collected during these projects, in partnership with GSC and University collaborators, will be critical for interpreting the setting of known

mineral deposits and to better assess the mineral-development potential of the region.

The Western Superior NATMAP project, now in its fourth year of operation, aims to provide a modern geoscience synthesis of northwestern Ontario and eastern Manitoba, using the combined resources of the federal, Ontario and Manitoba geological surveys. Key areas have been selected to address questions regarding relationships between Mesoarchean and Neoarchean sequences of the western Superior Province, through new mapping, geochemistry, geochronology and GIS-based compilation. Field and isotopic data on plutonic and supracrustal rocks from the northwestern Superior Province reveal a complex history involving three fault-bounded crustal terranes. These terranes record the collision between reworked 3.6 Ga crust in the north and a less than 3 Ga protocraton at 2.72–2.71 Ga.

A regional mapping program, begun in the Knee Lake area in 1997, continued to the northeast in the Gods Lake area in 2000. Geochronological studies yielded several important ages, indicating that the Oxford Lake Group volcanic subgroup has a 2720 Ma depositional age and that the Oxford Lake Group sedimentary subgroup has a broad range of detrital zircon ages ranging from 2711 to 3647 Ma. Geochronological data from the Knee Lake area date Hayes River Group volcanism at 2830 Ma. Folding of the Hayes River Group predates deposition of the coarsening-upward Opischikona clastic sedimentary package and iron-formation, the former containing 2937 to 2822 Ma detrital zircons. The Oxford Lake volcanic subgroup was emplaced at 2722 Ma in southern Knee Lake and was followed by deposition of fluvial sandstone with detrital zircons ranging from 2798 to 2707 Ma.

Mapping in the Ralph Anderson Lake greenstone belt, southeast of Thompson, was conducted as a follow-up to mapping carried out in the adjacent Max Lake area in 1999. The Ralph Anderson Lake belt extends from Logan Lake in the west to Aswapiswanan Lake in the east, a distance of about 50 km. The greenstone belt is essentially monoclinal, and consists mainly of arc-type mafic volcanic rocks, with subordinate felsic volcanic and sedimentary units. The widespread synvolcanic hydrothermal alteration prominent in the Max Lake area was also noted in this new map area. Stratabound sediment-related mineralization occurs at various stratigraphic levels within the mapped volcanic suites. Sulphide mineralization is also common in magnetiferous iron-formation units, which offer potential base-metal or precious-metal exploration targets. Sporadic pyritic zones that may represent prospects for economic mineralization also occur at the junction between the volcanic terrane and gabbro at the south margin of the greenstone belt. A gabbro-pyroxenite intrusion that extends through the central part of the Ralph Anderson Lake belt is currently being tested for possible PGE mineralization.

An M.Sc. study at the University of Waterloo focuses on the kinematic history of east-southeast-trending shear zones in the Pipestone Lake area of the Cross Lake greenstone belt. The south Pipestone Lake and north Pipestone Lake shear zones both underwent two main generations of shear deformation, the first associated with mylonite development under amphibolite-grade conditions and the second associated with shear under greenschist-grade conditions. Observations made during the structural study suggest that megacrystic anorthosite of the Pipestone Lake Anorthosite Complex intruded the basalt-dominated Pipestone Lake Group during or after an initial deformation event, suggesting that the proposed theory of comagmatism should be reconsidered.

Northwest Superior Boundary

An MGS- and NSERC-supported, integrated mapping, geochemistry and isotopic study of the Western Superior craton margin northeast of Thompson has been partnered with researchers from the University of Alberta. Over the past three years, this work has indicated that a re-interpretation of the location and nature of the boundary zone between the Archean Superior Province and the Paleoproterozoic Trans-Hudson Orogen is required.

In 1997, the first hints of ancient crust were discovered at Assean Lake. This discovery has led to extensive research in the Assean Lake area and the adjacent crustal domains of the Superior Boundary Zone. The Assean Lake ancient crust comprises a collage of Archean crustal segments, trending approximately 090–110°, that are overprinted by Neoarchean and Paleoproterozoic deformation and metamorphism trending 060°. The Assean Lake crustal complex is subdivided into 1) a southern, migmatitic, supracrustal series of quartz arenite, arkose, and metagreywacke gneiss, containing amphibolite and silicate-facies iron-formation, and 2) a central, orthogneiss-dominated crustal segment that intrudes 3) a northern package of supracrustal rocks dominated by mafic to intermediate metavolcanic rocks and greywacke gneiss. Combined Sm-Nd isotopic and U-Pb geochronological results, obtained using thermal-ionization mass spectrometry (TIMS) and sensitive high-resolution ion microprobe (SHRIMP), indicate that Assean Lake preserves Paleoarchean to Mesoarchean crust that underwent a complex and prolonged history spanning more than two billion years.

SOUTHEASTERN MANITOBA

Regional Mapping in Southeastern Manitoba

A collaborative GSC–MGS mapping program in the Black Island area of southeastern Lake Winnipeg was conducted under the auspices of the Western Superior NATMAP project. Initial reconnaissance studies were completed in 1999. The mapping was undertaken to characterize the nature of the boundary between two important tectonic elements of the western Superior Province, the North Caribou Terrane and the Uchi Subprovince. In the Black Island area, the margin of the ca. 3.0 Ga North Caribou Terrane is overlain by a quartzite–carbonate–iron-formation–komatiite sequence that records pre-2.92 Ga rifting of this protocontinent. Younger, Neoarchean volcanic rocks (mainly basalt and basaltic andesite) of the Uchi Subprovince were structurally juxtaposed against this margin in a ca. 2.72 Ga strike-slip regime. A succession of arkose and conglomerate containing detrital zircons as young as 2706 Ma was deposited in strike-slip basins during continued early oblique convergence. Waning deformation became focused in shear zones during subsequent transcurrent shear events. Elucidation of the complex structural history along the boundary of the Uchi Subprovince and North Caribou Terrane at the west end of the Rice Lake greenstone belt provides a much-improved geological context in which to conduct exploration activities.

PGE Studies

The Bird River Sill has historically been explored for chromite, Ni-Cu and platinum-group elements (PGE). The recent increases in both price and demand for PGE led to renewed interest in the Bird River Sill. Work this year on the Chrome property has defined a sulphide- and PGE-bearing layer hosted by highly deformed peridotite. This new PGE exploration prospect is in addition to two previously defined PGE-bearing layers associated with chromitite units. On the Page property, located about 2.5 km northeast of the Chrome property, new work has also defined a sulphide-bearing peridotite layer associated with chromitite units. This chromite- and PGE-bearing layer has a stratigraphic position analogous to those on the Chrome property.

Phytoremediation of Mine Tailings

A new project to establish the trace-element content of abandoned mine-site tailings, the potential for phytoremediation of these sites, and an assessment of the production of bio-ores through the identification of hyperaccumulator plant species has been initiated at the Central Manitoba (Au) mine site in southeastern Manitoba. The first phase of this project has established the trace-element characteristics of tailings at the mine site through multi-element analysis of samples collected from three, 1 m deep, hand-augured profiles. Results indicate that a partial

extraction at pH 2–3 liberates a wide range of base and precious metals from the tailings at the mine site, and that these metals provide the focus for the future development of a bio-ore. Ten plant species were planted on three experimental sites on mine tailings. The survival rate and amount of extracted base and precious metals should indicate plant species suitable for phytoremediation and phyto-extraction.

SOUTHERN AND CENTRAL MANITOBA

The MGS continues to promote exploration and development opportunities for specialty and nonmetallic minerals, both on the Precambrian Shield and in the Phanerozoic-covered portions of south-central Manitoba. The increasing evidence for post-Precambrian structural movements within the buried portions of the Superior Boundary Zone–Thompson Nickel Belt has provided a unifying theme for a number of features within the Phanerozoic section in west-central Manitoba.

Regional Investigations in the Interlake Region

The Camperville area of the Interlake region contains a large (65 km diameter) semicircular gravity anomaly of approximately 30 mGal, the most intense negative gravity anomaly in Manitoba. In the spring and summer of 2000, reconnaissance gravity surveys and follow-up detailed gravity and magnetic surveys were carried out along a line, from Camperville, Manitoba to the Saskatchewan border, that transected the gravity low. The Camperville anomaly is adjacent to the sub-Phanerozoic extension of the Thompson Nickel Belt and is apparently truncated on the northwest by a major fault defining the eastern edge of the TNB. Other recent studies have concluded that the Superior Boundary Zone has been active during the Paleozoic. Such late structures may have produced channelways for deep fluid migration from the buried Precambrian basement through the overlying Phanerozoic strata, and could have left their signature by depositing minerals at structural and lithological traps. Geochemical anomalies and the presence of sphalerite in drillholes penetrating Devonian formations in this region suggest that such fluid migration and mineralization may indeed have taken place during the Phanerozoic.

Saline spring waters have been sampled along the west shore of Lake Winnipegosis. Preliminary results suggest that Pleistocene glaciation had a significant impact on the regional flow system of the Paleozoic aquifers in the area.

Phanerozoic Stratigraphic Investigations

Stratigraphic investigations and drilling programs were carried out for various projects in south-central Manitoba. The Capital Region Project has been completed and is in the editing stage. H.R. McCabe's historical corehole data have been added to the Manitoba Stratigraphic Database. All brine springs are being tabulated for the Prairie-type Project. Two coreholes were drilled in the Dancing Point (Lake Winnipeg) area for stratigraphic purposes; one corehole was drilled proximal to the High Rock Lake structure; one corehole was drilled in the Wekusko area to test metallic veining in the Red River Formation; and one corehole was drilled in the Winnipegosis area for regional stratigraphy. A total of 457.8 m of drilling was conducted this year.

In 1999, the northern portion of the Capital Region Study was released as preliminary maps. In 2000, the study was extended to the south, and the maps and report are in the editing stage for the entire Capital Region Study, including the north and south halves. The Manitoba Stratigraphic Database (MSD) continues to be updated with the addition of historic tops to 437 wells.

Collaborative Projects in Southern Manitoba

As a contribution to the National Geoscience Mapping Program (NATMAP), the MGS and the GSC are co-operating in a program of geological mapping for the Winnipeg Region, with emphasis on engineering and environmental geology. Surficial mapping was completed in two phases for the area from 49° to 51°N and from 95° to 98°W. The current emphasis of the NATMAP and related MGS programs is on the completion of a three-dimensional (3D) digital geological model for all Phanerozoic sedimentary rocks and Quaternary deposits. This model will be completed first in the NATMAP area and subsequently for all of southern Manitoba. A number of data sets have been used to construct this 3D model: a digital elevation model (DEM), large lake bathymetry, offshore seismic surveys, surficial geology maps, lithological data for the Quaternary from water-well and other drillhole and geophysical databases, bedrock surface elevations, bedrock geology maps, Phanerozoic stratigraphic data and previous Phanerozoic models, and sub-Phanerozoic Precambrian geology maps. Completion of all project outputs is planned for March 2002.

The MGS and the GSC have initiated a multidisciplinary research program to reconstruct the pre-instrumental flood history of the Red River and to assess the importance of geological processes that may be increasing the flood hazard. The project is using a variety of geological and biological records to provide annual reconstructions of important hydroclimatic variables over the last several thousand years and provide a proxy record of high-magnitude floods in the Red River valley. The current tree-ring record, collected on a 100 km transect along the Red River, extends for 536 years from AD 1463 to 1999. This record contains anatomical signatures caused by Red River floods in 1950, 1852, 1826, 1811, 1778, 1747 and 1532. Complementary isotopic, chemical and ring-width records will be developed to evaluate the relationship between climatic conditions within the Red River drainage basin and the flood record of the Red River.

Aggregate Resource Studies

The MGS conducts aggregate-resource studies in southern Manitoba, work that contributes to the resource management and land-use planning decisions made by district planning boards. In 2000, 1:50 000 scale aggregate resource studies were conducted in the Rural Municipality of Cameron, covering eight townships in southwestern Manitoba near the town of Hartney. Gravel reserves (in two delta deposits) range from 3 m to more than 6 m in thickness and are composed of sand and gravel, with most of the reserves lying below the water table. Shale content of the gravel varies quite widely within the deposits; higher shale contents lower their economic value. Most of the aggregate produced comes from six active gravel pits and three shale quarries. The municipality is the greatest consumer of aggregate, mostly for road maintenance.

Land Use

The year 2000 marked the official end of the World Wildlife Fund Canada's Endangered Spaces Campaign. However, the provincial Protected Areas Initiative has been extended for three more years. This year, another 200 000 ha of the province were placed in protected status when Birch Island, Fisher Bay and Hudwin Lake Park reserves were designated in the spring. In addition, almost 40 000 ha of Wildlife Management Area lands were placed in protected status. Industry participation and consultation in the review of candidate sites for Manitoba's Network of Protected Areas provides Manitoba with increased certainty related to land-use issues.

NORTHERN MANITOBA

Industrial-Mineral Potential, Churchill

Precambrian and Phanerozoic bedrock outcrops and quarry locations in the Churchill area were inventoried during a brief visit to the region. Current production is mainly for aggregate and minor building stone. Although current production is considerably less than for most southern Manitoba localities, Churchill has potential for industrial-type

minerals because rail transportation costs would make alternatives prohibitively expensive. Future production may involve production of quartzite as a source of silica, rare elements from quartz veins and minor rock collecting. Unique fossils, such as trilobites, have been found along the shoreline of Hudson Bay. The largest recorded complete fossil of a trilobite ever found, a specimen 70 cm long by 30 cm wide, was collected from the Ordovician Churchill River Group on the tidal flat near Churchill in July 1998 by D. Rudkin of the Royal Ontario Museum.

GEOSCIENCE INFORMATION SERVICES

Compilation Maps

The Bedrock Geology Compilation Map Series (BGCMS) was initiated in 1985 to provide uniform 1:250 000 geological compilation maps for the entire province, initially focusing on areas of high mineral endowment and/or high mineral potential. Digitization of all existing manually drafted BGCMS maps is almost complete, and map production is now 100% digital, using ArcInfo™ GIS technology. There are now 29 BGCMS maps covering more than half of the province, including the entire Churchill Province, core areas of the Superior Province, and the capitol region of southern Manitoba. Those parts of Manitoba not yet covered by the BGCMS include areas of mainly unexposed Phanerozoic bedrock and portions of the Superior Province with little or no bedrock exposure or no supracrustal rocks.

Web-Enabled GIS Products

Web-enabled geographic information system (GIS) technology is a tool used to present geospatial data on the Internet. This technology provides the Internet client with twenty-four hour access to our department's information. To access the database, an Internet connection and a JAVA-enabled web browser such as Microsoft Internet Explorer™ or Netscape Navigator™ is required. The website has a user-friendly GIS style, with tools that allow query and analysis of the most current data sets relevant to the mining and mineral exploration community. The 'GIS Map Gallery', which is the access point for information on mineral claims, assessment files, orthophotography and geological databases, can be found at:

http://www.gov.mb.ca/em/geoscience/gis/gis-index.html.

CLIENT SERVICES

As a result of a departmental reorganization in August 2000, responsibility for the production and sale of technical reports reverted to the Geological Survey. The Client Services group is also responsible for the Library, a Divisional resource that is much used by the public.

February 2000 marked the first time the *Report of Activities* was posted to the Internet, on the departmental web site. Clients are able to download PDF files (for individual reports) from the web site at no cost and print pages identical to those in the hard-copy report. In the first six months following the posting of the *Report of Activities 1999*, there have been almost 2200 PDF files downloaded by clients, making this one of the most popular methods of report distribution. Further enhancements to the Geological Survey web site in 2000 included the introduction of web pages for a diverse range of topics:

- Paleofloods in the Red River Basin
- Preliminary exploration database for platinum-group elements in Manitoba (OF2000-5)
- Digital elevation model of southern Manitoba
- GIS map gallery

REGIONAL OFFICES

The Flin Flon regional office continues to supply a variety of services to the mineral exploration and mining communities active in the Flin Flon–Snow Lake region, including recording new mineral claims, maintaining an up-to-date library of provincial claim maps, dealing with claim status inquiries and accepting assessment-work submissions. Progress was made in the ongoing mines documentation project for the Flin Flon–Snow Lake region. Efforts are currently being directed toward completing several outstanding reports, including three volumes of the *Mineral Deposit Series*. The present staff in the Flin Flon office includes a Regional Geologist, a Resident Geologist and a Recording Clerk.

Staff in the Thompson regional office respond to a range of inquiries including regional geology, potential of mineral properties and mineral identifications. Thompson staff geologists are currently involved in a comprehensive multidisciplinary re-evaluation of the geology and mineral wealth of the Lynn Lake greenstone belt, the Thompson Nickel Belt and PGE potential in the Bird River sill. The Thompson office provides Mining Recording services to the community, including data on the status and registration of claims, access to and sales of maps and reports, and access to electronic databases.

DEPARTMENTAL INCENTIVE PROGRAMS

Departmental initiatives that foster a positive business climate for exploration and mining include the introduction of more favourable taxation policies, continuation of the Mineral Exploration Assistance Program (MEAP), the Prospectors Assistance Program (MPAP), and the new Specialty Minerals Incentive Program (SMIP), aimed at promoting development in the industrial-minerals sector.

Mineral Exploration Assistance Program (MEAP)

The purpose of the Mineral Exploration Assistance Program is to increase exploration and stimulate activities that may lead to the development of new mines in Manitoba. The program makes $2.75 million available in each of three fiscal years, beginning in 1999–2000.

MEAP will provide assistance for up to 25% of approved eligible exploration expenses, to a maximum of $300 000 per company per fiscal year. The program offers up to 35% of approved eligible expenses, to a maximum of $400 000 per company per fiscal year, for proposed projects in remote northern areas and in the Lynn Lake–Leaf Rapids area. Priority is given to grassroots or preliminary exploration (i.e. exploration not associated with a producing mine).

Since the inception of MEAP in the fall of 1995, 331 applications for assistance have been submitted. A total of $16.8 million in financial assistance has been allocated to 269 projects proposed by 76 companies. To date, the total amount of assistance issued is $8.2 million. Exploration expenditures in Manitoba are reported to be $36.0 million per year.

Prospectors Assistance Program (MPAP)

The Manitoba Prospectors Assistance Program offers grants to prospectors exploring on either their own properties or on open Crown mineral land in Manitoba. The program provides 50% of the eligible expenditures incurred by prospectors, up to $7500 per applicant per year. To qualify, applicants must hold a Manitoba Prospecting Licence, be experienced and self-employed, and have a prospecting project within Manitoba on mineral dispositions held by the applicant or on open Crown mineral land. Eligible expenditures include wages, subsistence, transportation, operating supplies, rental of equipment, assays, diamond drilling, and contractor and/or consulting services.

Applications for summer projects must be submitted between April 1 and May 15 of the year in which the project is to be carried out, or between November 1 and December 15 for winter projects. Since its inception in 1992, the program has processed 285 applications for assistance and paid out $744 148 in assistance funding to prospectors.

Specialty Minerals Incentive Program (SMIP)

The Specialty Minerals Incentive Program is designed to assist companies and individuals to evaluate the economic potential of existing specialty mineral deposits and market those resources more effectively. Funding of up to $500 000 was identified for SMIP, beginning August 3, 1999. The program will make $250 000 available for each of the 1999–2000 and 2000–2001 fiscal years.

SMIP provides financial assistance for up to 30% of approved eligible expenditures, to a maximum of $50 000 per applicant per fiscal year. The eligible commodities targeted by SMIP include, but are not limited to, amber, bentonite, kaolin, other clay minerals, gypsum, limestone, silica sand, coal, salt, shale, peat, dimensional stone, spodumene, chromite, talc, soapstone, feldspar, titanium, magnesium, beryl, ilmenite, potash, lignite and hematite. Minerals that are not eligible under SMIP include aggregates as they pertain to construction, base and precious metals, and oil and natural gas. Eligible expenditures include bulk sampling, definition drilling, geochemical analysis and assaying, feasibility and market studies (as they pertain to the resource), and testing of samples.

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Virtually the entire technical and nontechnical staff of the Manitoba Geological Survey has played a role in the production of the *Report of Activities 2000*. The meticulous work of field geologists, lab technicians, expeditors, students, cartographers and publication staff is gratefully acknowledged. We also thank external reader B. Davie of RnD Technical for careful editing and helpful suggestions for improving geologists' manuscripts.

GS-1 GEOCHEMISTRY AND METAL MOBILITY DURING SILICIFICATION OF THE 1.89 GA WELCH LAKE PRIMITIVE ARC BASALT, SNOW LAKE ARC ASSEMBLAGE, SNOW LAKE

by M. Surka[1] and A.H. Bailes

Surka, M. and Bailes, A.H. 2000: Geochemistry and metal mobility during silicification of the 1.89 Ga Welch Lake primitive arc basalt, Snow Lake arc assemblage, Snow Lake; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 9-11.

INTRODUCTION

During the 2000 field season, a small outcrop of silicified Welch Lake basalt was mapped at 1:10 scale as part of a B.Sc. thesis at the University of Manitoba. The objective of the study is to establish whether the observed alteration is simply a result of redistribution of elements at the outcrop scale or the result of exchange of elements at a larger scale. If the silicification is part of a regional-scale exchange of elements, then a further objective is to establish the amounts of metal mobilized during alteration, in order to assess the potential for the directly overlying sulphidic sedimentary rocks of the 'Foot-Mud horizon' to contain significant base-metal volcanogenic massive sulphide (VMS) mineralization.

REGIONAL SETTING

The study area is located at the west end of the Paleoproterozoic Flin Flon greenstone belt, in a sequence of isotopically juvenile, ca. 1.89 Ga volcanic rocks, more than 6 km thick, belonging to the Snow Lake arc assemblage (Fig. GS-1-1). The Snow Lake arc assemblage, which contains eight Cu-Zn VMS mines in the Snow Lake area, comprises volcanic rocks that display an upward change from relatively more primitive arc, to relatively more mature arc, to ocean-floor basalt (Bailes and Galley, 1996; Fig. GS-1-2). Both the primitive and mature arc sequences are bimodal in composition, comprising basalt, rhyolite and large subvolcanic tonalite plutons. Volcanogenic massive sulphide deposits in both sequences are hosted by rhyolite and spatially associated with regionally extensive zones of altered volcanic rocks (Fig. GS-1-2). The Snow Lake primitive arc, which is composed of low-Ti refractory basalt lavas (Welch Lake formation), boninite lavas, and isotopically juvenile felsic flows and tonalite plutons, has been interpreted by Bailes and Galley (1999) to be the result of high-temperature hydrous melting of refractory mantle sources in an extensional and/or proto-arc environment.

A prominent zone of silicification, the subject of this study, has affected the upper 500 m of the Welch Lake basalt at the top of the primitive arc sequence (Fig. GS-1-2, -3). This zone has a strike length of more than 20 km and is directly overlain by sulphidic sedimentary rocks (Foot-Mud horizon) containing disseminated pyrite and pyrrhotite. Although low in base-metal content, the Foot-Mud horizon represents a significant episode of sulphide deposition in the Snow Lake region.

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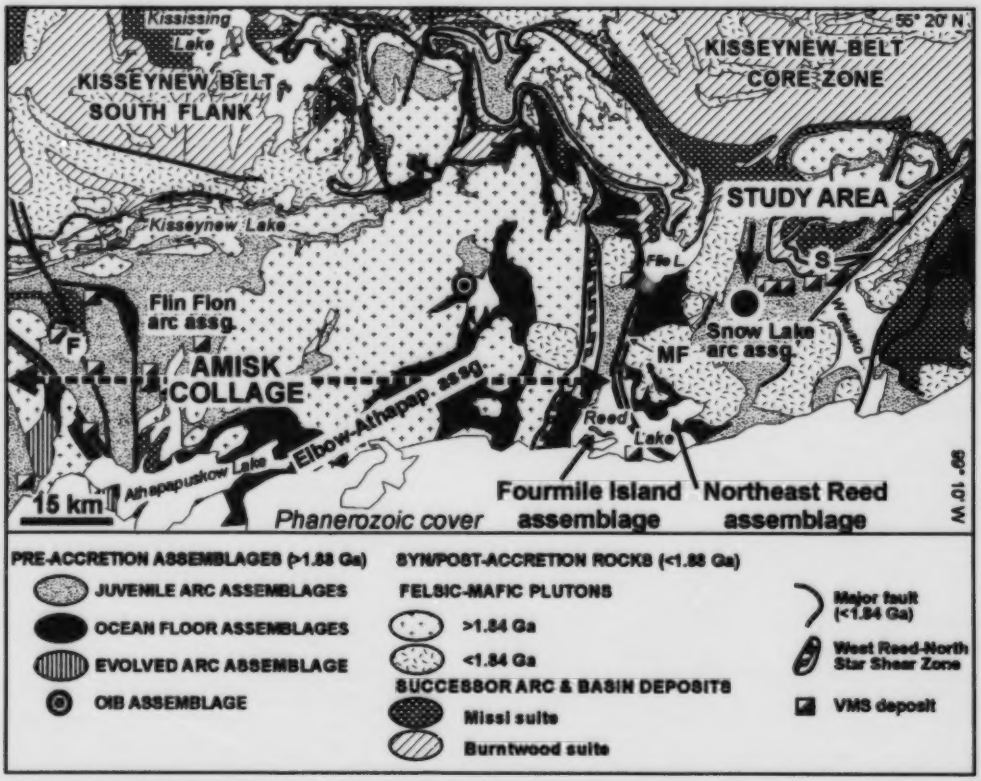

Figure GS-1-1: Simplified geological map of the central and eastern portion of the Flin Flon Belt, showing major tectonostratigraphic assemblages and plutons, and locations of mined VMS deposits. Abbreviations: F, Flin Flon; S, Snow Lake; MF, Morton Lake Fault Zone. Black dot shows location of the study area in the Snow Lake arc assemblage at the east end of the Flin Flon Belt.

Figure GS-1-2: Schematic cross-section of the Snow Lake arc assemblage, showing distribution of the primitive, mature and arc-rift volcanic rocks, and the overprinting alteration zones. Note the location of the study area at the top of the primitive arc sequence within a regional-scale, conformable zone of silicification.

FIELD MAPPING AND SAMPLING PROCEDURES

The mapping and sampling of the outcrop of silicified Welch Lake basalt was undertaken as follows:

1) The outcrop was carefully cleaned and a 1 m control grid established so that features could be accurately mapped and documented.
2) Primary features in the basalt flow (pillows, interpillow hyaloclastite, amygdales, thermal contraction cracks) were depicted on the map to evaluate their control on alteration.
3) Individual alteration facies, which are presently characterized by distinctive metamorphic mineral assemblages produced during almandine-amphibolite facies regional metamorphism at ca 1.81 Ga, were identified and then mapped.
4) Using a rock saw, samples were collected from each of the alteration facies, including separate sample sets from pillows and interpillow yaloclastite. Late fractures and veins, as well as weathered surfaces and amygdales, were avoided during sample collection.
5) Samples from the pillow core, pillow margin and interpillow hyaloclastite of a relatively unaltered basalt flow in the same section were collected for comparison with the altered basalt samples.
6) The study outcrop and adjacent outcrops were examined for features that could be used to establish the timing of the silicification event. Although the alteration is considered to be synvolcanic, it could also have formed later (e.g. during regional metamorphism).

DATA AND SAMPLE PROCESSING

The alteration facies on the study outcrop will be digitized so that the areal extent and distribution of the various alteration facies can be readily and accurately established. This, in conjunction with geochemistry of samples from the various alteration facies, will enable us to establish the redistribution of elements during alteration.

Polished thin sections from each of the alteration facies will be examined and analyzed by electron microprobe to provide detailed characterization of the mineral assemblages in each of the alteration facies. Geochemical analyses by ICP–MS will be used to identify the geochemical signature of each of the alteration facies.

a)

Figure GS-1-3: a) Study outcrop, showing silicification affecting a series of pillows, with author for scale. b) Individual pillow, showing the variability in intensity of alteration; each zone is characterized by different mineral assemblages, which permit accurate identification and mapping of the alteration facies.

b)

Mass-balance chemical comparison of samples from the silicified basalt and from relatively unaltered equivalents will provide the basis for determining element redistribution in the various alteration facies. These data, combined with the areal extent of the various alteration facies, will be used to determine whether the silicification 1) resulted from redistribution of elements in a 'closed' system (i.e. outcrop-scale), or 2) was externally introduced during a larger, regional-scale alteration event. The mass-balance change of metals during the silification of the study outcrop will indicate whether or not the alteration process was capable of contributing metals to the directly overlying Foot-Mud horizon.

SUMMARY

Mapping of an outcrop of silicified basalt at 1:10 scale was undertaken as part of a B.Sc. thesis at the University of Manitoba. This thesis study, which will be completed by the spring of 2001, comprises detailed mapping and sampling of the outcrop, and comprehensive mineralogical and geochemical studies of the collected samples. It is designed to fully characterize the nature of the alteration, and to provide an understanding of the silicification event that is more complete than that resulting from regional mapping and alteration studies conducted by Bailes and Galley (1992, 1996), Skirrow and Franklin (1994) and Bailes et al. (1996).

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ACTIVITIES OF THE FLIN FLON REGIONAL OFFICE

by T. H. Heine and D. E. Prouse

Heine, T.H. and Prouse, D.E. 2000: Activities of the Flin Flon regional office; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 12.

SUMMARY

The Flin Flon regional office continued to supply a variety of services to the mineral exploration and mining communities active in the Flin Flon–Snow Lake region, including recording new mineral claims, maintaining an up-to-date library of provincial claim maps, dealing with claim-status inquiries and accepting assessment-work submissions. Progress was made in the ongoing mines-documentation project for the Flin Flon–Snow Lake region. Efforts are currently being directed toward completing several outstanding reports, including three volumes of the *Mineral Deposit Series*. The present staff includes a Regional Geologist (Heine), a Resident Geologist (Prouse), and a Mining Recorder or Recording Clerk (position to be filled).

MINE DOCUMENTATION PROJECT

Based on metal content per unit area, the Snow Lake–Flin Flon–Hanson Lake (Saskatchewan) greenstone belt is the most prolific Paleoproterozoic greenstone belt in the world (Franklin et al., 1995). Recent investigations, conducted mainly under the auspices of the NATMAP Shield Margin and LITHOPROBE Trans-Hudson Orogen projects (*see* Syme et al., 1998 for references), have made great strides in elucidating the geological framework of this region. Despite the advances made by these studies, detailed investigations of the base- and precious-metals deposits are limited to concise summaries, such as the Manitoba Industry, Trade and Mines *Mineral Deposit Series* reports, or descriptive papers published over a long period in a variety of journals. The detailed geology of these deposits, particularly relative to the mine workings, is generally poorly documented. Neither has a collection of rock and ore samples representative of the deposits been assembled as a resource for future reference.

This project was initiated with the documentation of the Photo Lake deposit which, at that time, was in the final stages of exploitation. Many of the underground workings were still accessible, so a representative suite of rock and ore samples could be assembled relatively easily. In addition, Hudson Bay Mining and Smelting Co. Ltd. (HBM&S) and Hudson Bay Exploration and Development Co. Ltd. (HBED) agreed to support the project and gave full access to their digital plans, sections and databases of this deposit. As a result of this invaluable co-operation, we now have a suite of reference samples, a number of complete exploration drillholes, a geochemical database for samples collected from the exploration holes, and a set of plans and sections showing the detailed geology of the mine. The sample suite and core are permanently stored indoors at the Manitoba Industry, Trade and Mines core-viewing facility, located at the reclaimed Centennial mine site near Bakers Narrows, approximately 20 km southeast of Flin Flon.

This project has now been expanded to include some of the other deposits in the Chisel Lake area, including the Ghost, Lost and Chisel mines, and the Chisel North orebody. Accurately located samples from the underground workings can no longer be obtained from the first three deposits, so this phase of the program will rely on sampling of drill core. All the mine drawings, with the exception of Chisel North and the Chisel Lake open pit, will need to be digitized, and this work is currently in progress.

It is planned to issue the data sets and interpretation for each mine, possibly as an open file report or a CD-ROM volume. The combination of samples and data will provide a valuable resource for future investigations in this prolific belt.

REPORTS

Several reports were completed, or are in the final stages of preparation. The *Mineral Deposit Series* report for NTS area 63N/01 (Nightingale Lake) went to press in June. The report for 63K/10 (Iskwasum Lake) has been sent for review, and the one for 63K/15 (Elbow Lake) is in the final stages of preparation. With the publication of these reports, coverage of the Flin Flon–Snow Lake region by the *Mineral Deposit Series* will be complete.

An open file report describing the geology of the Persian Lake area (NTS 63K/13) is in the final stages of preparation. The most recent mapping of this area dates from the mid-1940s and fails to mention the extensive rhyolitic sequence exposed here. This report will be completed by year end.

The activities of the mining and mineral-exploration communities within Manitoba were actively monitored, and summary reports prepared for both internal use and external publication.

MINING RECORDING

Mining recording and associated functions continued to be supplied through this office by the Mining Recording Clerk. An up-to-date collection of claim maps for the province is maintained for public reference, new claims are recorded and assessment reports are accepted by our office. When requested, reports summarizing the status and history of specific claims are prepared for clients.

EXPLORATION ASSISTANCE

The main function of the Flin Flon office is to provide assistance to the mining and mineral-exploration communities, and all clients interested in the geology of the region. Clients are given access through this office to the Precambrian Drill Core Libraries, located in The Pas, Thompson and Lynn Lake. An exterior, secure core-storage repository is available adjacent to the Centennial core-viewing facility for clients who wish to protect their core from vandalism or loss.

Several nonexploration industry clients availed themselves of our services, including several educational institutions and nongeological organizations.

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GS-3 THOMPSON NICKEL BELT PROJECT: PROGRESS ON A NEW COMPILATION MAP OF THE THOMPSON NICKEL BELT (PARTS OF NTS 63F, 63G, 63J, 63O AND 63P)

by J.J. Macek, C.R. McGregor and H.V. Zwanzig

Macek, J.J., McGregor, C.R. and Zwanzig, H.V. 2000: Thompson Nickel Belt project: progress on a new compilation map of the Thompson Nickel Belt (parts of NTS 63F, 63G, 63J, 63O and 63P); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 13-15.

SUMMARY

Geological compilation of NTS sheets 63F/08, 63F/09, 63G/14 and 63J/03 was completed during last winter as part of an ongoing compilation map program for the Thompson Nickel Belt (TNB). The program is a collaborative project involving the Manitoba Geological Survey, Inco Ltd., Falconbridge Limited and Hudson Bay Exploration and Development Company Limited (HBED).

During the summer, diamond-drill core from 317 holes, drilled in a broad area around Wabowden, was re-examined. Thirty per cent of these drillholes contain Ospwagan Group rocks. This is notable, as no Ospwagan Group rocks were shown on the 1972 compilation map of this area. It is well known that, in the northern part of the TNB, the Ospwagan Group metasedimentary sequence hosts large nickel deposits, including the Thompson Mine deposits.

Although only one-third of the available core was examined this summer, major lithological trends will be compiled onto maps from the newly obtained data. Relogging of additional drillholes will be undertaken next summer to further refine the compilation map.

COMPILATION MAP PROGRESS

Work on the new compilation map of the TNB, sub-Paleozoic portion (Macek et al., 1999a) continued during the winter months by extending the compilation (Fig. GS-3-1) to:

1) the extreme southwest (NTS sheets 63F/08 and 63F/09); and
2) the central part (NTS sheets 63G/14 and 63J/03), filling the gap between Falconbridge Limited and HBED permit data.

Compilation work in the southern part of the exposed TNB around Wabowden (Fig. GS-3-1, dotted pattern) was delayed until re-examination of the diamond-drill core, originally stored at the Bucko mine site (Macek et al., 1999b), had begun. Relogging of this core was initiated this summer. Out of the 15 000 boxes of core, 5000 were relogged, providing new regional lithological information on 317 diamond-drill holes. Most of these holes are randomly distributed thorough the area, which straddles Phillips, Halfway, Setting, Bowden, Rock Island, Resting, Conlin, Gormley, Key and Muningwari lakes.

RESULTS

A variety of ultramafic rocks, such as serpentinized dunite, metaperidotite, metapyroxenite, porphyroblastic metapicrite and derived ultramafic schists, was identified in 57% of the examined holes (Fig. GS-3-2a). Ospwagan Group metasedimentary rocks were identified in 30% of the relogged holes (Fig. GS-3-2b). The frequency with which all Ospwagan Group formations were intersected is shown in Figure GS-3-2c, where M, T, P, S and B stand for Manasan, Thompson, Pipe, Setting and Bah formations, respectively.

The abundance of Pipe Formation rocks encountered in the drillholes (57%) is mainly a result of 'target drilling' of strong magnetic anomalies associated with electromagnetic conductors. In the TNB, magnetic anomalies are usually caused by ultramafic rocks and magnetiferous silicate-facies iron-formation. Those conductors caused by graphite-enriched, sulphide-facies iron-formation occur in the Pipe Formation (Macek and Bleeker, 1989; Bleeker, 1990).

CONCLUSIONS

The abundance of Ospwagan Group rocks in the re-examined drillcore is quite significant, because the previous compilation map of the TNB (Coats et al., 1972) does not show any occurrences of the Ospwagan Group metasedimentary sequence (unit 5) in the Wabowden area. Previously, the most southerly occurrence of Ospwagan Group rocks was recorded on Brostrom Lake, just north of Sasagiu Rapids. The re-examination of the retrieved core, initiated this summer, will substantially contribute to the accuracy of the new compilation map of the TNB. Although two-thirds of the available core have yet to be examined, major lithological trends have emerged and will be added to the compilation map during the coming winter.

ACKNOWLEDGMENTS

The authors gratefully acknowledge the generous logistical and personnel support of Falconbridge Limited, including free access to proprietary data; without this assistance, the compilation work would not have been possible.

We also thank our summer assistants, C. Lettley, F. McGregor, N. Cates, S. Johnston, K. Salzsauler and J. Thackeray, for their valuable help during work that was often conducted under difficult conditions.

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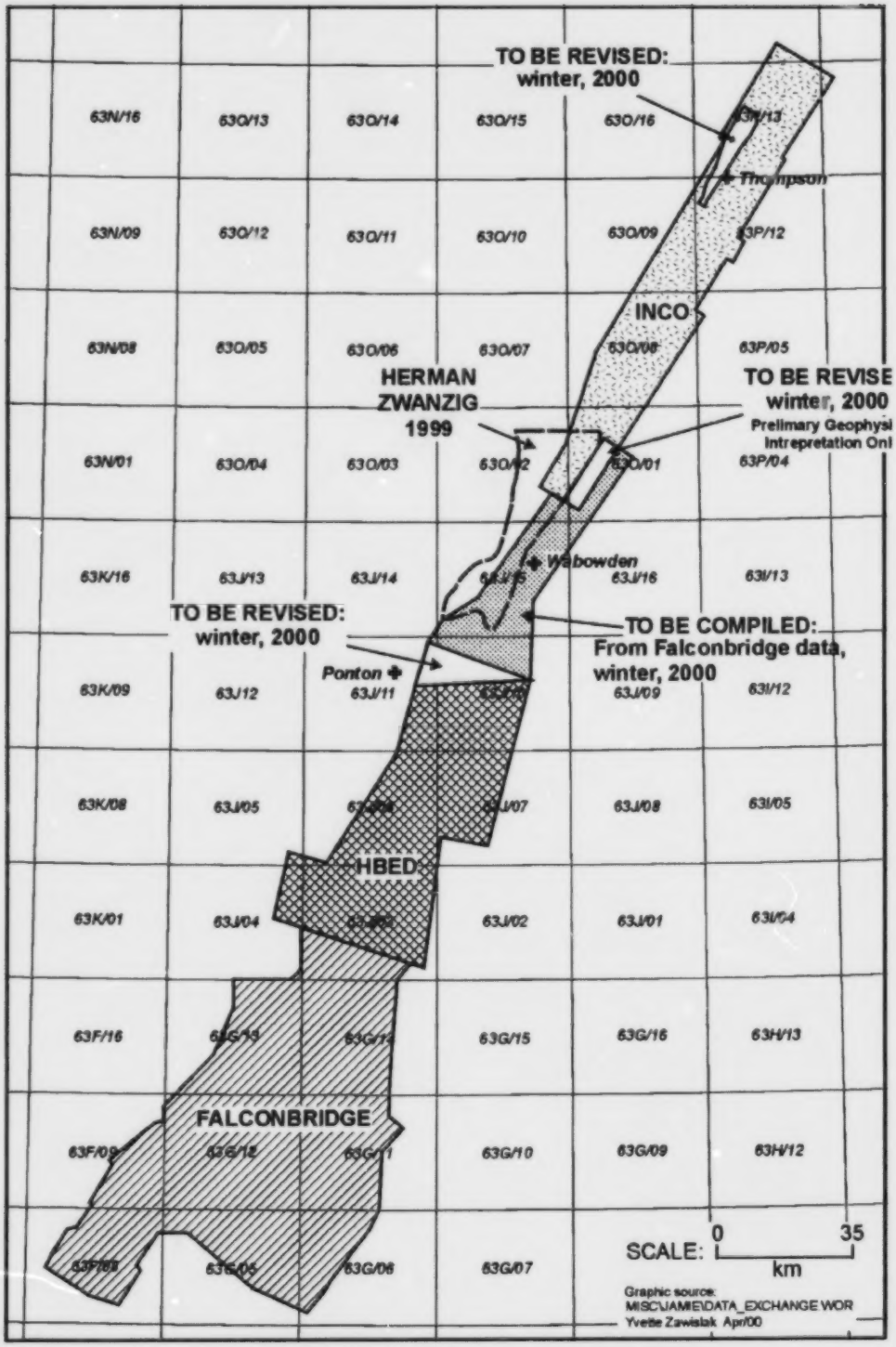

Figure GS-3-1: Status of the Thompson Nickel Belt compilation, showing boundaries of the geological data contributed by all partners.

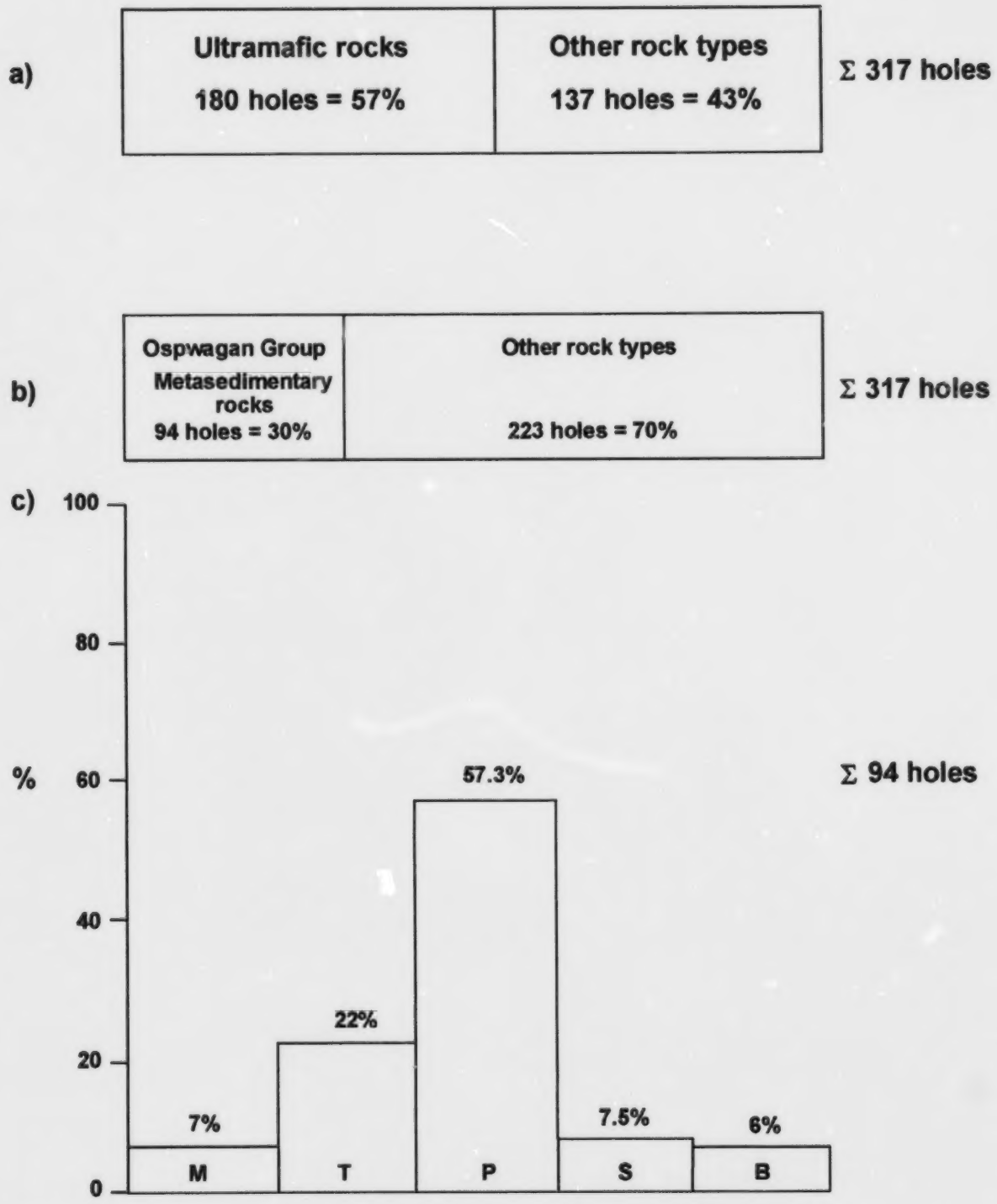

Figure GS-3-2: Main rock types encountered in the re-examined diamond-drill holes in the region around Wabowden.

GS-4

THOMPSON NICKEL BELT PROJECT: PETROGRAPHIC AND CHEMICAL CHARACTERIZATION OF MAJOR ROCK TYPES (PARTS OF NTS 63J, 63O AND 63P)

by C.R. McGregor

McGregor, C.R. 2000: Thompson Nickel Belt Project: petrographic and chemical characterization of major rock types (parts of NTS 63J, 63O and 63P); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 16-17.

SUMMARY

Petrographic studies of the Ospwagan Group supracrustal rocks in the Thompson Nickel Belt (McGregor, 1997) were initiated to provide detailed mineralogical data for the Ospwagan Group. The work was later expanded (McGregor, 1998) to include other rock types that are in close proximity to, or in contact with, the Ospwagan Group rocks, namely the Kisseynew gneiss and the reworked Archean gneiss, respectively. In addition, geochemical data for all studied rock types were compiled. Microprobe analyses of principal minerals were completed. Evaluation of data and preparation of the report are in progress.

RECENT INVESTIGATIONS

In the winter of 1999, modal analyses of Ospwagan Group metasedimentary rocks, Archean basement gneiss and Kisseynew metasedimentary rocks were supplemented by 74 partial modal analyses published in previous geological reports (Fig. GS-4-1). Mineralogical trends for the various rock types were established.

Ninety-eight microprobe analyses of principal minerals from the Ospwagan Group sequence exposed in the Pipe II open pit mine were conducted in order to determine their proper mineralogy. Many of the

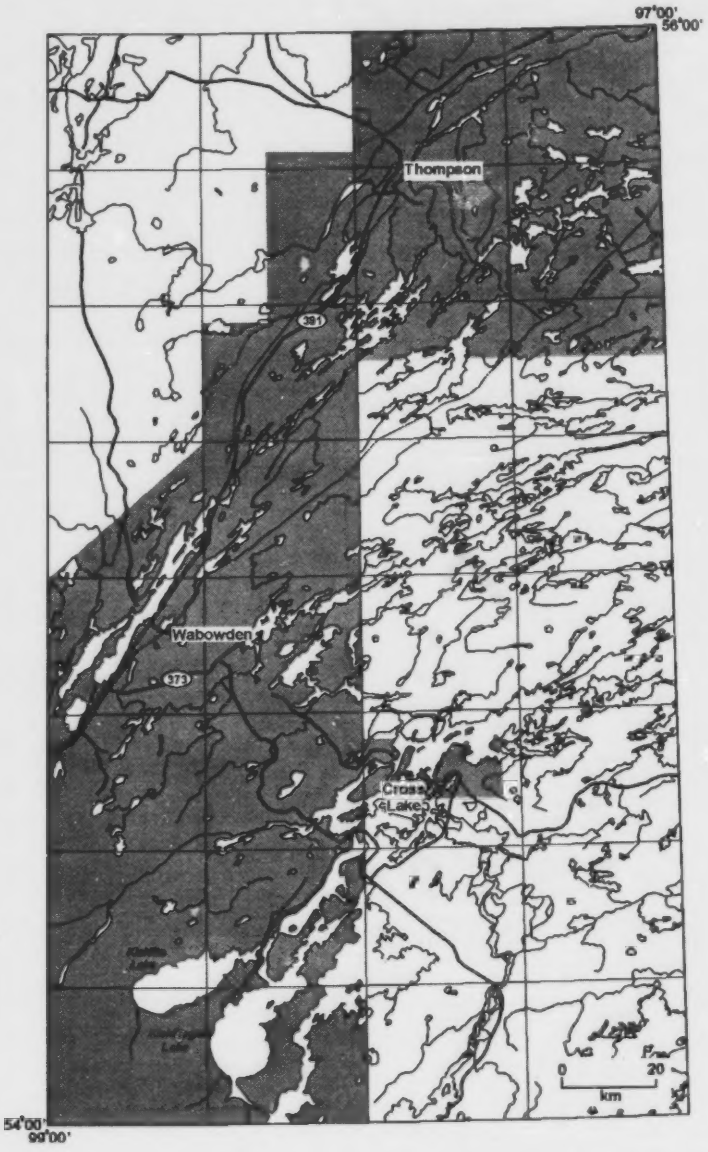

Figure GS-4-1: Area of all partial modal analyses done by previous workers (between 1954 and 1978).

selected minerals are mafic and occur in iron–formation units. All data gathered will be evaluated and incorporated into the Ospwagan Group report, currently in preparation.

OTHER INVOLVEMENT

Relogging of diamond-drill core from 5000 core boxes (out of 15 000 core boxes), representing 317 drillholes, was completed during the 2000 field season (*see* Macek et al., GS-3, this volume).

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McGregor, C.R. 1997: Thompson Nickel Belt project: Petrographic and chemical characterization of major lithologies (parts of NTS 63J, 63O and 63P); *in* Report of Activities 1997, Manitoba Energy and Mines, Geological Services, p. 102.

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GS-5

MYSTERY LAKE HIGH-STRAIN ZONES: SINISTRAL AND DEXTRAL, EAST-SIDE-UP DISPLACEMENTS, THOMPSON NICKEL BELT

by H.V. Zwanzig

Zwanzig, H.V. 2000: Mystery Lake high-strain zones: sinistral and dextral, east-side-up displacements, Thompson Nickel Belt; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 18-23.

SUMMARY

A regional syncline, which was traced northeast along Mystery Lake in previous mapping, has a core of Ospwagan Group metasedimentary and metavolcanic rocks with large ultramafic intrusions. The Archean basement gneiss on the limbs of the syncline contains high-strain zones that feature kinematic indicators interpreted, in this report, to be the result of northwest shortening with sinistral east-side-up displacement and, locally, less clearly defined dextral east-side-up displacement. With respect to a structural history proposed by Bleeker (1990a, b) and modified by Zwanzig (1998), the large fold and the high-strain zones developed mainly during D_3 sinistral transpression. The Mystery Lake data suggest that northwest compression and vertical extension were dominant but that mega-boudinage may have caused local concentration of the shear component and rare reversals in the horizontal shear component. The results do not support a model involving only dextral transpression (A. Potrel, D. Peck, N. Machado and C. Wegleitner, work in progress, 2000). However, they are consistent with the structural style of undisrupted, long sinuous folds that has been identified during preparation of a compilation map of the Thompson Nickel Belt (Macek and McGregor, 1998).

INTRODUCTION

The complex, long-lived structural evolution of the Thompson Nickel Belt (TNB) has lead to a variety of tectonometamorphic interpretations. Bleeker (1990a, b) proposed early nappe tectonics, followed by intrusion of 1.88 Ga dykes and late (ca. 1.77 Ga) sinistral transpression. Zwanzig (1998, 1999) presented a revision to the model of Bleeker (1990a, b), citing from his work in the Setting Lake region 1) the apparent absence of early, large-scale, stratigraphic repetitions in the Ospwagan Group; and 2) the presence of large recumbent folds, formed ca. 1.81 Ga, during the time of peak metamorphism of the Kisseynew gneiss. Potrel et al. (work in progress, 2000) have proposed that the TNB structure formed as a result of dextral transpression alone. Whereas final interpretation of the structural fabrics at Setting Lake awaits the completion of the regional compilation map (Macek and McGregor, 1998), a known pattern of major map units at Mystery Lake provides the critical framework for a kinematic analysis in the northern part of the TNB.

Mesoscopic structural data were collected on Mystery Lake over a period of 10 days in spring, before the cresting Burntwood River had flooded the critical outcrops. The Mystery Lake shoreline has a varied relief and provides clean subhorizontal exposures, cleavage surfaces and south faces, produced by glacial abrasion and plucking. Consequently, the various planar and linear fabric elements were measured, and kinematic indicators were commonly recorded from observations on the three principal surfaces. In this report, the structural measurements from various types and ages of fabrics are analyzed and related to structures that are several kilometres long. These are related, in turn, to the regional structure of the TNB. A special effort was made to analyze the movement pattern around the large ultramafic bodies hidden under the lake (Coats et al., 1972).

STRATIGRAPHIC AND STRUCTURAL FRAMEWORK

Mystery Lake is underlain by steeply dipping sedimentary and volcanic rocks of the Ospwagan Group, exposed on several islands and, locally, on the shore (Fig. GS-5-1). Archean basement gneiss occurs on much of the east and west shores and forms small outrops on the adjacent high ground. The gneiss is overlain by the lower part of the Ospwagan Group (mainly Manasan Formation quartzite and Pipe Formation semipelite). The upper formations in the Ospwagan Group (Setting quartzite with thin pelite, and Bah Lake mafic flows and sills) occur on the central islands. Consequently, stratigraphic facing is toward the centre of the lake, inward from both sides, defining a large syncline. A west-facing angular unconformity at the base of the Ospwagan Group in the northeast corner of the lake is consistent with this interpretation (Fig. GS-5-2; *see* 'U' on Fig. GS-5-1). The large, differentiated, North Mystery Lake ultramafic intrusion, which occupies much of the east limb of the syncline, also faces west (H.V. Zwanzig, work in progress, 2000). The Mystery Lake granodiorite batholith lies directly west of the syncline.

The dip of the layering is subvertical everywhere, indicating that the fold is isoclinal. The supracrustal units pinch and swell along strike; slightly downward-diverging dips on the limbs of the syncline are taken as an indication that the Ospwagan Group also pinches and swells on a large scale about a shallow-plunging axis in the core of the fold.

Small-scale folding is prominent in strongly layered (stromatic) basement gneiss and, locally, in well bedded Ospwagan Group sedimentary rocks. Curvilinear hinges occur in some areas but they curve through less than 90°, which suggests an origin by superposed folding, as at Setting Lake (Zwanzig, 1998), rather than sheath folding.

The structural style and fabric elements at Mystery Lake are similar to those along strike at Setting Lake (Zwanzig, 1998). The deformational history is therefore also considered to be the same, and the main Proterozoic high-temperature foliation is interpreted as composite S_2-S_3 schistosity with coplanar elements that are distinguished in F_3 fold hinges at Setting Lake (Ducharme and Zwanzig, 1999). At Mystery Lake, the basement rocks show a strong early gneissosity (S_A) and stromatic layering, both of which are assumed to be Archean because they are not present in the Ospwagan Group. The Proterozoic biotite schistosity (S_2-S_3) is generally coplanar with S_A but cuts it in some of the early (?Archean) fold hinges. In much of the Mystery Lake area, as elsewhere in the TNB, the intense late fabric development (S_3, L_3) obliterated or transposed the earlier fabrics. Weakly developed crenulation in S_2-S_3 has a variable plunge and an uncertain origin. The S_4 foliation, containing fine-grained muscovite and chlorite or carbonate, is associated with late shear zones.

HIGH-STRAIN ZONES

High-strain zones occur 1) close to the contact between the Ospwagan Group and the basement gneiss, 2) along the east margin of the Mystery Lake granodiorite, and 3) at the west margin of a uniform granitoid gneiss that is part of the basement on the east side of the 'northeast area' (Fig. GS-5-1). These zones feature strong planar fabrics and moderately well developed lineations. They comprise high-temperature protomylonitic to mylonitic foliations and stretching lineations that overprint the Archean gneissosity (S_A) in the basement and the schistosity (S_2) in the supracrustal rocks. They are thus interpreted to represent mainly S_3. Lens-shaped bodies, such as boudins, pods of leucosome (Fig. GS-5-3) and porphyroclasts within the highly planar fabric, are predominantly symmetric and indicate a large component of coaxial strain: northwest compression, at a high angle to the regional strike (Fig. GS-5-1). Most lineations plunge subvertically down dip. They are

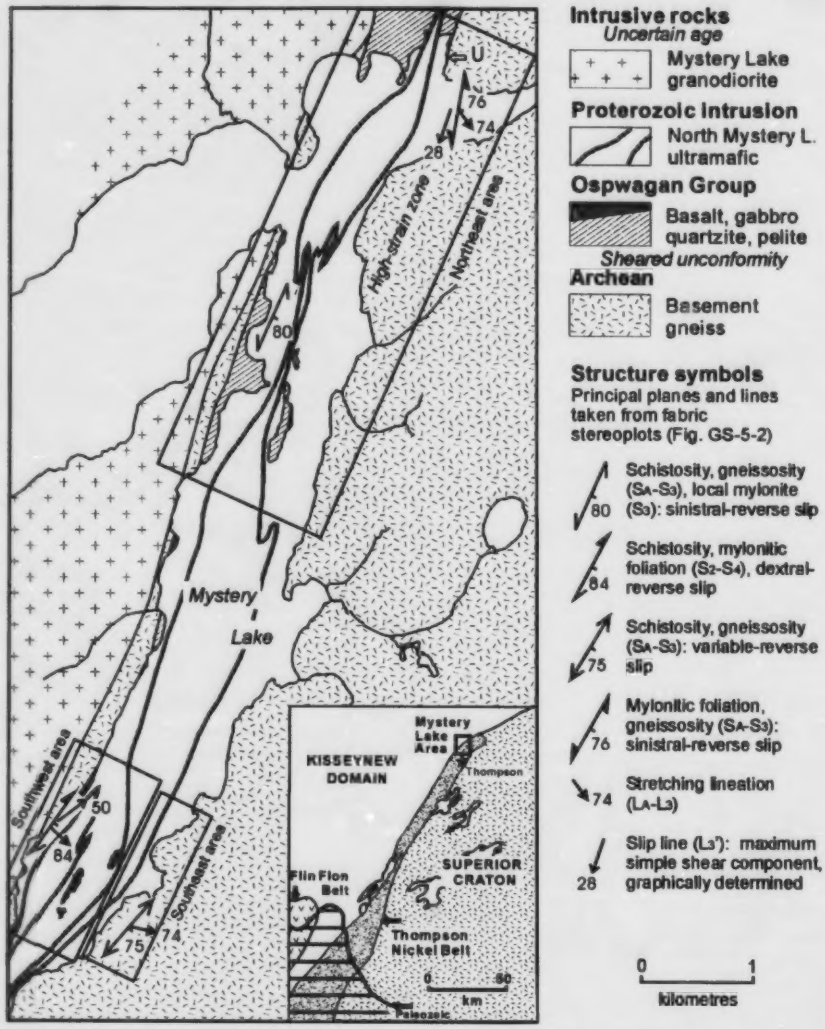

Figure GS-5-1: Geological sketch map of the Mystery Lake area with the outline of the North Mystery Lake ultramafic intrusion (after Coats et al., 1972), and showing the principal planar and linear fabric orientations (for three structurally distinctive areas) as determined on Figure GS-5-8.

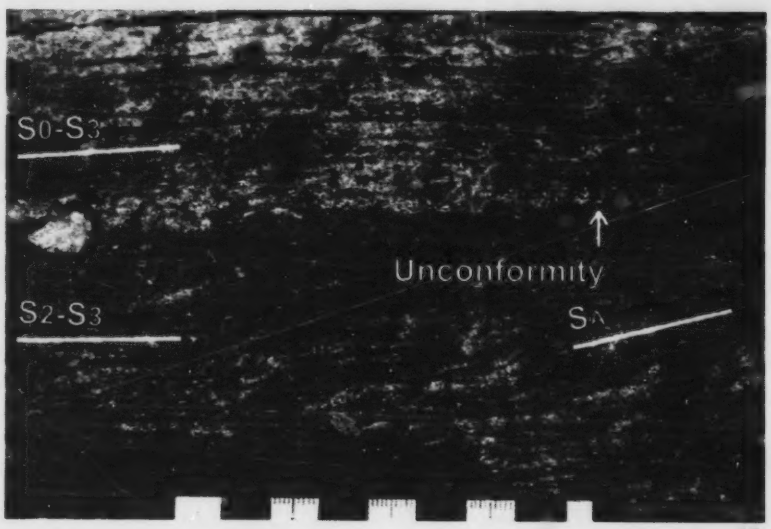

Figure GS-5-2: Unconformity at the base of the Ospwagan Group, showing the angular relationship between 1) Archean foliation (S_A) in the basement gneiss, which is overprinted by weak Proterozoic schistosity (S_2-S_3), and 2) thin bedding with coplanar schistosity (S_0-S_3) in the overlying quartzite.

Figure GS-5-3: Lenses and folded layers of granitoid leucosome in layered (stromatic) basement gneiss. The slight dextral asymmetry of the larger sigmoidal pods (above the tape) are analogous to the structural style of the North Mystery Lake ultramafic intrusion (see Fig. GS-5-1). Tape is 10 cm long.

dominated by stretching lineations that indicate a high degree of coaxial vertical extension, because asymmetrical structures are absent or subdued on vertical rock faces perpendicular to the foliation.

The most prominent mylonite zone occurs in the Archean granitoid gneiss in the eastern part of the 'northeast area'. However, the highest strain in the basement may be recorded in the strongly gneissic to schistose, multicomponent stromatic rocks along the shores of Mystery Lake in the 'southeast area' and 'southwest area' (Fig. GS-5-1). This unit contains a sedimentary component and is similar to Ospwagan Group semipelite injected by granitic veins. Consequently, the origin of some of the strongly foliated rocks on the west side of the lake is uncertain. Zones with even higher strain may occur in the incompetent pelite units under the lake. In contrast, well preserved, weakly foliated to massive rocks occur in the interior of the Mystery Lake granodiorite and on the central islands. Well preserved basalt, ultramafic rock and quartzite appear to occupy low-strain domains that are probably lens-shaped, though not fully exposed. The high-strain zones around them are probably anastomosing but only exposed along the shore. The strain gradient on the margin of the well preserved domains implies that the local strain was noncoaxial (Hanmer and Passchier, 1991).

Noncoaxial shear-sense indicators occur sporadically along the west shore of Mystery Lake and consistently in the high-strain zone in the northeast area. The mylonite contains S-C fabrics (S_3-C_3; Fig. GS-5-4), and more discrete shear bands (C_3'), which formed sequentially during progressive deformation (Hanmer and Passchier, 1991), occur in the protomylonite gneiss (Fig. GS-5-5, -6). Asymmetrical structures, such as winged sigma porphyroclasts and back-rotated boudins, generally show the same sense of shear as the planar structures. Rolled (delta) porphyroclasts are rare or absent. Discrete narrow shear zones (Fig. GS-5-7) and small faults containing ultramylonite and pseudotachylite do not all have a consistent sense of shear. They are considered to be relatively young (S_4), but their relationship to the main high-strain zones is uncertain. Late shear zones on Setting Lake have greenschist-facies retrograde mineral assemblages in amphibolite-grade rocks, a relationship that will be tested in the rocks from Mystery Lake.

MESOSCOPIC FABRIC ANALYSIS

All planar fabrics in the Mystery Lake area strike northeast to north and are vertical or have a steep easterly dip (Fig. GS-5-8). Overturning of layering and axial surfaces of folds, or vergence, is consistently to the northwest. Overall, the area shows a slight curvature, with the more northerly strikes being in the northeast. The distribution of subvertical foliation and steeply plunging stretching lineations defines strong central maxima, typical of strongly preferred orientations of L-S fabrics with random variations (Fig. GS-5-8). This fabric data and the generally poor development of asymmetrical shear-sense indicators suggest that all rock units at Mystery Lake have undergone northwest compression and vertical stretching. The dispersion of poles and lineations is interpreted to be the result of local variations in the movement pattern around competent bodies. This is particularly evident in the 'southwest area', where boudins of gabbro and elongate lenses of leucogranite are present.

Asymmetrical fabric elements, which are indicative of noncoaxial strain, developed locally around Mystery Lake, but measurements have been concentrated in areas where they occur. The study area is subdivided into three domains (northeast, southeast and southwest areas) based on consistent changes in the shear sense from sinistral to dextral and to mixed (Fig. GS-5-1, -8). All of these domains show a component of reverse displacement (i.e. southeast-side-up). Independent measurements of the S and C fabric elements, as well as schistosity and mylonitic foliation with more discrete shear bands (C'), were taken on the same parts of the outcrops, and these data were used to approximate the vector of maximum shear strain (slip line). The C-planes are generally not exposed, and fine striations (Lin and Williams, 1992) were not observed. The 'slip' lines in Figure GS-5-8 are given by the normal to the S-C intersection in the plane of the main foliation (either S or C). They are determined with reasonable accuracy only in the 'northeast area', where eight pairs of measurements provided a tight cluster of lines that plunge south-southwest at about 30° (*see* lower right diagram in Fig. GS-5-8). An approximate direction of slip is suggested by the symmetries on several outcrops, where strongly asymmetrical sinistral indicators on a horizontal surface are matched by weakly asymmetrical but consistent indicators for reverse displacement on subvertical exposures normal to the foliation. All relevant data indicate that the slip line plunges southwest throughout the 'northeast area'.

In the 'southwest area', the sense of slip appears to be dextral and a single pair of surfaces suggests a moderately northeast-plunging slip line. Again, this was predicted by comparing horizontal and steep faces, but here the asymmetry is weak as viewed on the horizontal plane. The calculated vector is almost normal to that in the northeast. The greater dispersion of stretching lineations in this area (*see* upper left diagram in Fig. GS-5-8) may be related to the local variations in ductility (*see* above). This may reflect the partitioning of the simple shear component of strain into narrow zones in which the stretching lineations approach the slip line, whereas the shear-zone-normal shortening is reflected in the general steep plunge of the lineations (Lin et al., 1998). This is consistent with the spread of lineations in Figure GS-5-8. Future work will include a thorough search for and sampling of such zones.

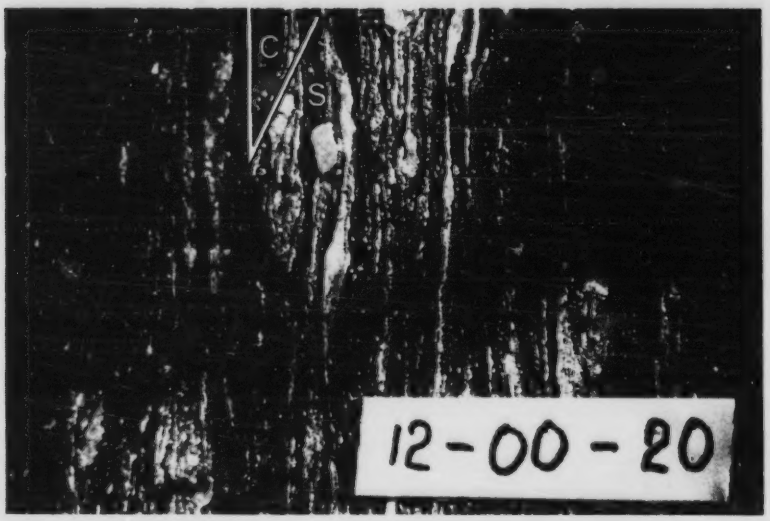

Figure GS-5-4: Layered mylonitic basement gneiss with S-C fabric indicating east- side-up displacement; view looking northeast. Its protolith is interpreted to have been like the rock in Figure GS-5-3.

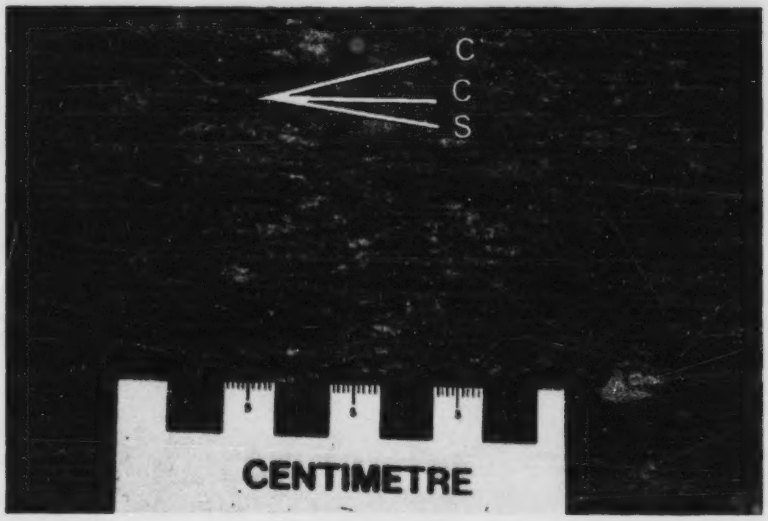

Figure GS-5-5: Archean granitoid gneiss overprinted by Proterozoic foliation (S-C) with shear bands (C'), as seen in plan view in the protomylonite zone in the 'northeast area' on Figure GS-5-1.

In the 'southeast area', dextral indicators tend to occur on shoreline promontories, close to the North Mystery Lake ultramafic intrusion. Sinistral indicators occur in the bays and suggest a reversal in the flow regime away from the intrusion. One outcrop featured indicators suggesting both shear senses. However, as on the southwest shore, the asymmetry is not pronounced on horizontal outcrops.

DISCUSSION

The shear-related fabrics are interpreted to be the result of the D_3 and D_4 deformation, which occurred during the development and tightening of the F_3 major syncline at Mystery Lake. The strong, vertical, coeval stretching is consistent with the northwest compression that resulted in the folding. A component of noncoaxial strain produced oblique, sinistral, southeast-side-up displacement throughout the northern half of the lake. Clearly, the competent southeast basement block was uplifted and moved northeast, starting to override the northwest bock that contained the Mystery Lake granodiorite. The incompetent pelitic gneiss in the core of the large syncline at Mystery Lake acted as a ductile zone, probably with partitioning of oblique simple shear into zones around low-strain domains.

Relatively shallow-plunging oblique displacement with vertical stretching, as is described in this report, is a hallmark of transpression zones. However, several factors mitigate against a large component of transcurrent slip. These include: 1) a preponderance of structures indicating vertical extension and northwest compression, and the apparent absence of zones with horizontal striation; 2) generally weak development of C surfaces; 3) the lack of rotation of the stretching lineation into the calculated slip line in the largest high-strain zone in the northeast; and 4) a change in the movement direction around a large competent body that acted as a megaboudin and produced a reversal in the direction of the horizontal flow component from sinistral to dextral. Considerable transcurrent slip may be partitioned into pelites that are unexposed at Mystery Lake, but this is not apparent in pelite exposed in the Grass River linear northeast of Setting Lake.

An area of particular interest is located southeast Mystery Lake, where a reversal in horizontal flow suggests that a southward-widening basement wedge was expelled toward the southwest and upward. A ramp at the south side of this block would be a basement overthrust. A recumbent fold at the south side of the block would resemble a small, south-verging basement nappe. The area is accessible and warrants further mapping to test for such structures. A structural model that may be worthwhile to consider in the TNB is one of protracted northwest compression with a small regional component of oblique, sinistral, southeast-side-up displacement.

An important conclusion of this study of kinematic indicators on Mystery Lake is that the main part of the TNB is unlikely to represent a crustal-scale transcurrent zone. This is consistent with the structural

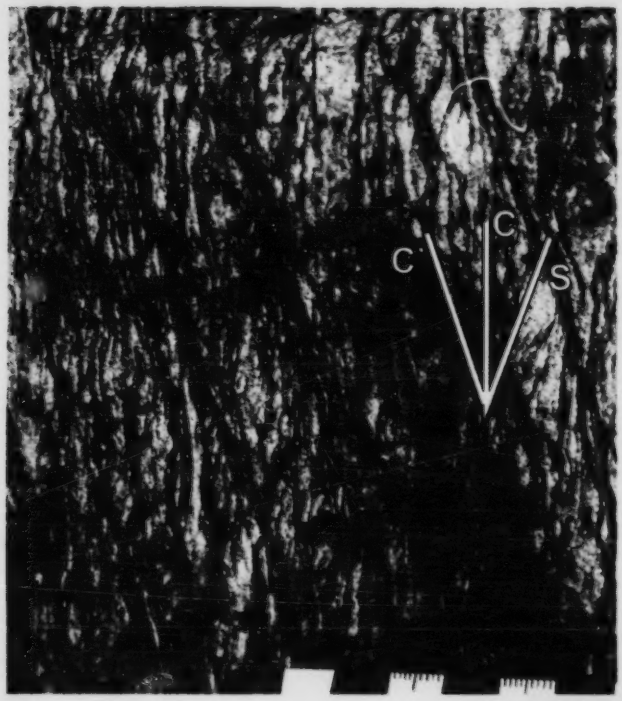

Figure GS-5-6: Archean granitoid gneiss overprinted by Proterozoic foliation (S-C) with shear bands (C'); view looking north, normal to and directly below the view in Figure GS-5-5.

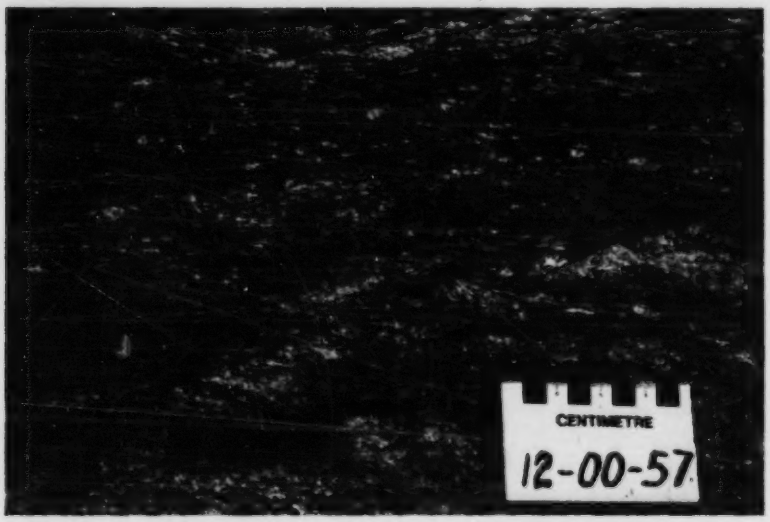

Figure GS-5-7: Archean granitoid gneiss, preserved in the lower right and cut by a late sinistral shear zone (?S_4). The abrupt increase in Proterozoic strain in the shear zone requires partitioning of considerable simple shear into the bounding fault.

style indicated by the regional map pattern (Coats, 1972; Macek and McGregor, 1998). Map units are continuous along strike, as well as across the strike of high-strain zones where these terminate or become highly curved. The units in the Ospwagan Group, not the high-strain zones, control the location of ultramafic intrusions and their associated nickel deposits in the TNB.

ACKNOWLEDGMENTS

I gratefully acknowledge Inco Ltd. for the access to and prompt help in obtaining their confidential geological map of Mystery Lake, which allowed me to put the structural data into perspective. I thank Mike Surka for his efficient and cheerful assistance in the field. Although I take full responsibility for the conclusions, the guidance of Chris Beaumont-Smith in the structural analysis and the scientific editing were invaluable.

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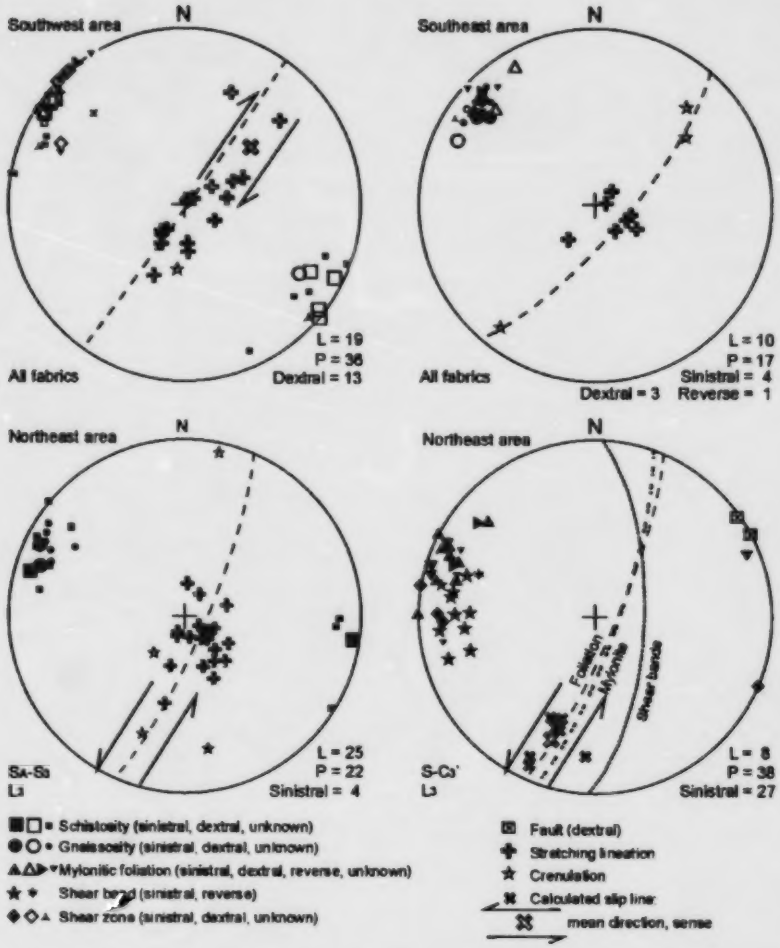

Figure GS-5-8: Equal-area stereoplots of various types and ages of fabric elements shown for three different areas on Mystery Lake. The 'slip lines' were determined as the normals to the S-C or C' intersections projected onto the main foliation (S or C). (This assumes rotation of earlier shear bands into S and may introduce a small error in the azimuth of the 'slip line'.)

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PETROLOGICAL INVESTIGATIONS AT SIPIWESK LAKE (PART OF NTS 63J/16)

by C.D. Lettley[1]

Lettley, C.D. 2000: Petrographic investigations at Sipiwesk Lake (part of NTS 63J/16); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 24-25.

SUMMARY

Five days were spent at western Sipiwesk Lake examining the metamorphic transition between the Archean Pikwitonei Granulite Domain and the Paleoproterozoic, amphibolite-grade, Thompson Nickel Belt terrane. Field work focused on collecting samples from layers of leucocratic, mesocratic and melanocratic rocks that could be reliably traced across this well defined metamorphic boundary. In addition, a series of samples was collected from a well exposed, layered mafic-ultramafic complex of the Pikwitonei Domain to study metamorphic reaction textures at granulite grade. Results of these studies will be reported in a B.Sc. thesis, which is now in progress.

GEOLOGY AND SAMPLING

Hudsonian Front

The Pikwitonei granulites (Mezger et al., 1990) at western Sipiwesk Lake are characterized by isoclinally folded layers of leucocratic and mesocratic granulite gneiss, invaded by layers of 'M2' mobilizate (Hubregtse, 1980), which consists of plagioclase+quartz±hornblende±orthopyroxene. The granulite also hosts rare ultramafic boudins and is cut by Molson dykes (Scoates and Macek, 1978; Ducharme, 1999).

Effects of the Trans-Hudson orogeny extend into western Sipiwesk Lake. Progressing west from the unaltered granulite in the eastern part of the area, the first visible effects of the Hudsonian overprint are zones of retrogression spatially associated with millimetre- to centimetre-scale veins filled with quartz and followed by pegmatite. Individual lithological layers can locally be traced across a relatively sharp boundary of retrogression (Fig. GS-6-1), allowing for the sampling of both retrogressed and pristine rocks from the same layer. Farther west, toward the Thompson Nickel Belt, metamorphic retrogression is accompanied by a tectonic overprint and by increase in the abundance of subhorizontal pink pegmatite veins. Samples of specific lithological assemblages were gathered from localities exhibiting differing degrees of retrogression and deformation.

Layered Mafic-Ultramafic Complex

A layered mafic-ultramafic complex, at the mouth of the Nelson River's west channel, was examined and sampled in detail to study metamorphic mineral assemblages and derive original mineralogical compositions of individual layers. Within the mafic portion of the complex, assumed to be metagabbro, layering is accentuated by the presence of abundant garnet in some layers, and its absence in others. Locally, the garnet can be seen to be developed from orthopyroxene, usually in very coarse texture, by a macroscopic reaction involving orthopyroxene, plagioclase and quartz.

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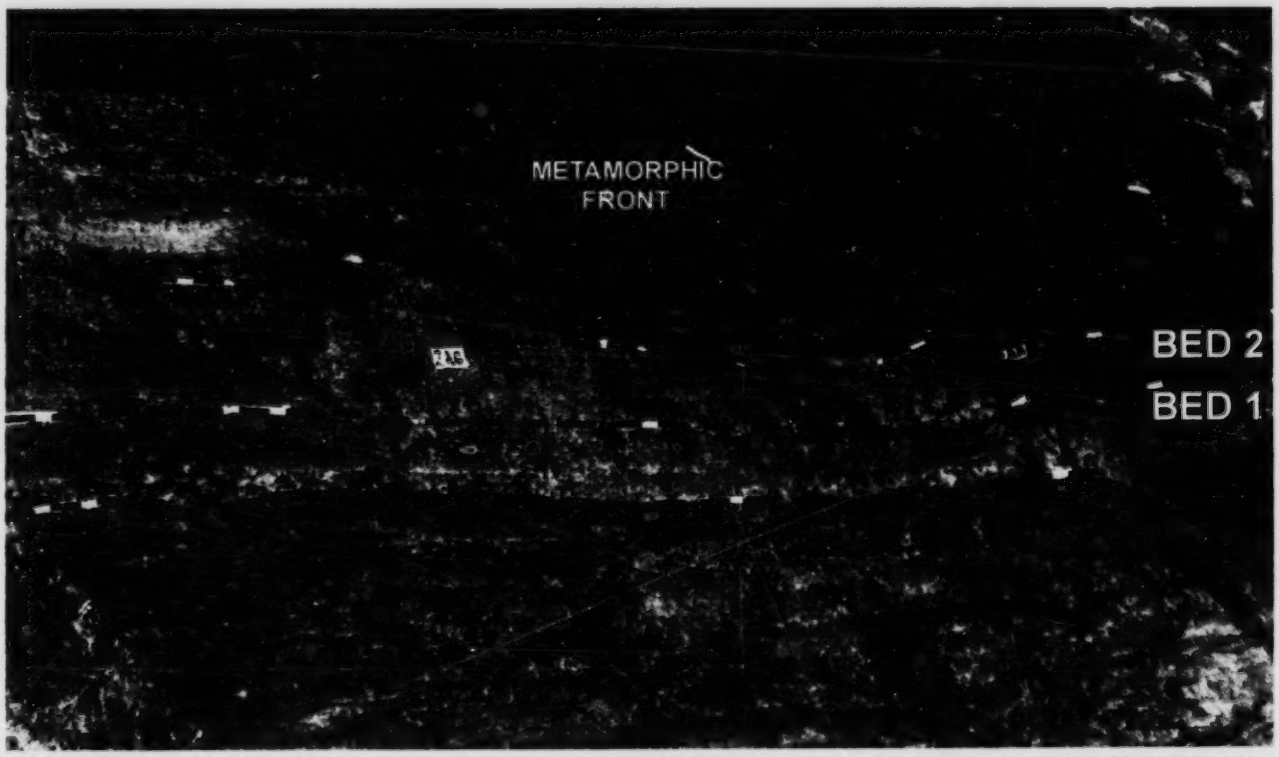

Figure GS-6-1: Zone of retrogression (upper right) cuts obliquely through layering in granulite. Individual layers (highlighted by white tape) can be traced from unaltered granulite into retrograde amphibolite. Note compass for scale.

In one location, well developed coronas (Fig. GS-6-2) consisting of orthopyroxene+quartz+garnet+plagioclase were sampled in order to characterize the garnet-producing reactions. These are the first coronas found in mafic rocks on Sipiwesk Lake. No coronas in the ultramafic part of the complex, analogous to those described by Macek (1989), were observed.

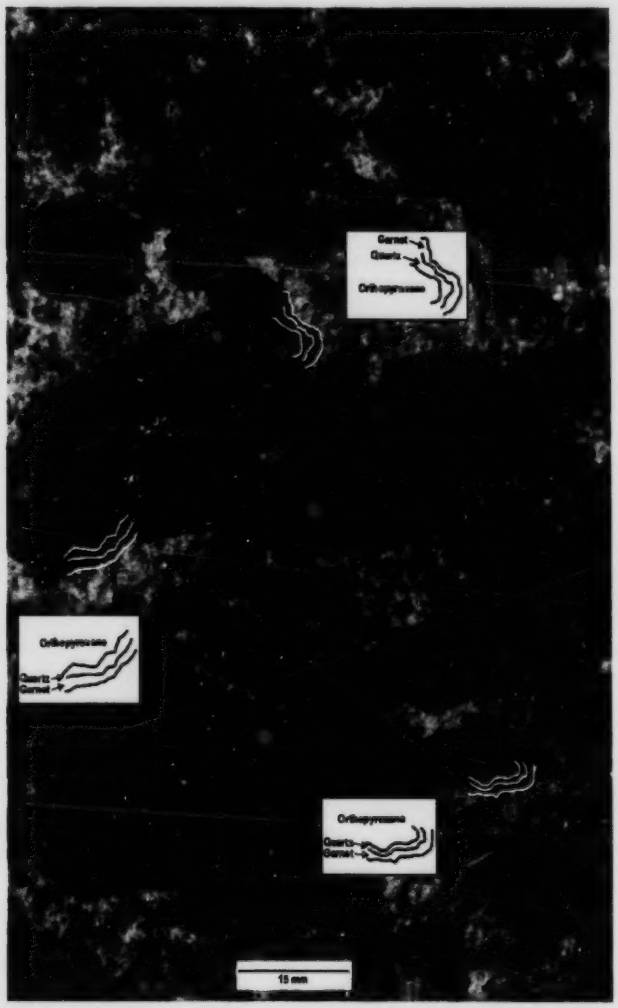

Figure GS-6-2: Coronas from layered mafic body. Orthopyroxene is rimmed by quartz and garnet. Image is approximately 1.5 times actual size.

FUTURE WORK

Over the course of the 2000-2001 academic year, microscopic, microprobe and chemical analyses of the samples will be performed. An analysis of the results will be conducted and research findings will be presented in fulfilment of B.Sc. requirements.

ACKNOWLEDGMENTS

I would like to thank Kristin Salzsauler (University of Manitoba) for her assistance in the field and Robert Springer (Brandon University) for his input into the study. I would also like to thank the Manitoba Industry, Trade and Mines rock laboratory technicians for their assistance in sample preparation. My greatest appreciation goes to Josef Macek and Cathy McGregor for sharing with me their experience, resources and, most importantly, their enthusiasm.

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GS-7

SPLIT LAKE BLOCK OF THE SUPERIOR BOUNDARY ZONE: PETROLOGICAL, PALEOMAGNETIC AND GEOBAROMETRIC DATA FOR THE CA. 1818 MA FOX LAKE PLUTON (NTS 64A/1, /8)[1]

by M.J. Harris[2], D.T.A. Symons[2], W.H. Blackburn[3], D.C. Peck[4] and A. Turek[2]

Harris, M.J., Symons, D.T.A., Blackburn, W.H., Peck, D.C. and Turek, A. 2000: Split Lake Block of the Superior Boundary Zone: petrological, paleomagnetic and geobarometric data for the ca. 1818 Ma Fox Lake pluton (NTS 64A/1, /8); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 26-37.

SUMMARY

The ca. 1818 Ma Fox Lake pluton of the Split Lake Block on the northwestern margin of the Superior craton is one of several rock units collected and analyzed paleomagnetically to describe the tectonic motions for the craton during the waning stages of the Trans-Hudson Orogen. The results suggest that this pluton was emplaced ca. 1850 Ma and later deformed ca. 1810 Ma. Detailed petrological observations and preliminary pressure-temperature data are also presented for the pluton and host rocks.

INTRODUCTION

Paleomagnetic directions from 20 rock units in the Trans-Hudson Orogen (THO) of Manitoba and northern Saskatchewan (Fig. GS-7-1) have been measured in the Paleomagnetic Laboratory at the University of Windsor. These data have yielded several important results:

- There is a classical paleomagnetic signature for continent-continent collision of the THO terranes ca. 1830 Ma.
- A stillstand in the apparent polar wander path (APWP) may exist from ca. 1830 to 1810 Ma.

[1] LITHOPROBE Report #1173

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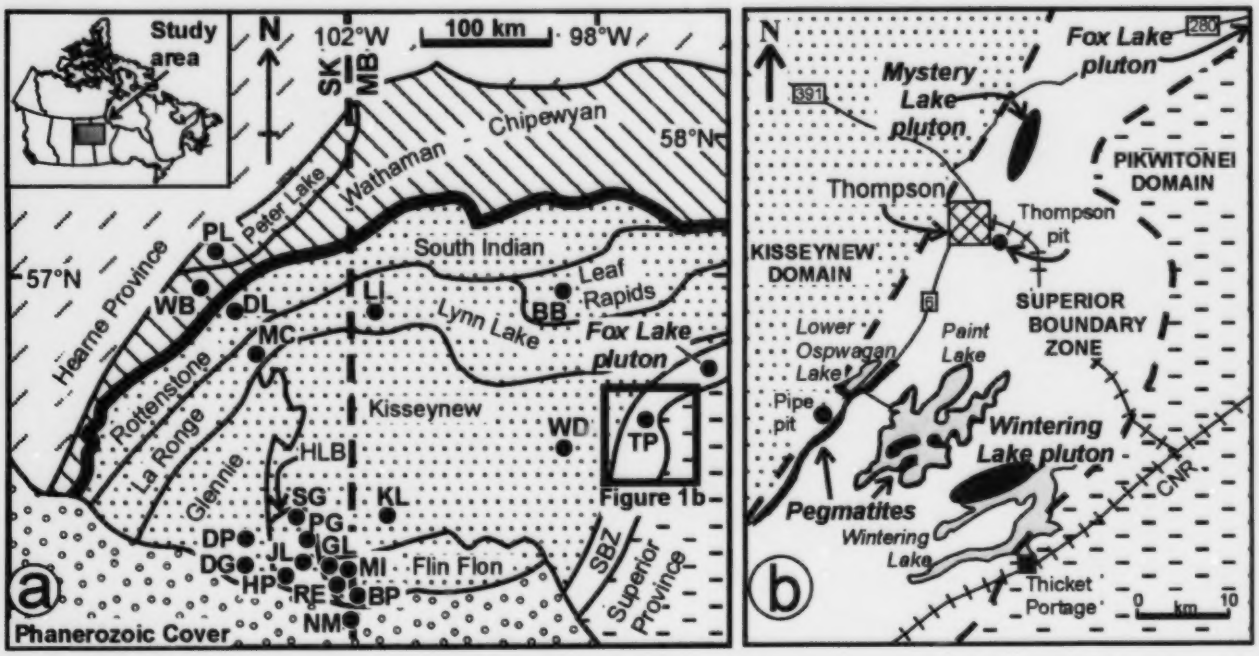


Figure GS-7-1: a) Tectonic elements of the Trans-Hudson Orogen (stippled region) and location of the Fox Lake pluton (Harris et al., 1999a, b). Heavy line represents the suture between the Trans-Hudson Orogen and the Slave, Rae and Hearne provinces. Abbreviations: HLB, Hanson Lake Block; MB, Manitoba; SBZ, Superior Boundary Zone; SK, Saskatchewan. Dots show paleomagnetic collection locations: BB, Baldock Batholith (Symons et al., 1996b); BP, Boot Lake–Phantom Lake complex (Symons and MacKay, 1999); DG, Deschambault Lake granite (Symons et al., 2000); DL, Davin Lake granite (Symons et al., 1996a); DP,– Deschambault Lake pegmatites (Symons et al., 2000); GL, Granite Lake pluton (abandoned); HP, Hanson Lake pluton (Gala et al., 1994); JL, Jan Lake granite (Gala et al., 1995); KL, Kississing Lake dome (Symons and Harris, 1998); LL, Lynn Lake gabbro (Dunsmore and Symons, 1990); MC, Macoun Lake granite (Symons et al., 1994); MI, Missi conglomerate (Symons and Lewchuk, 1994); NM, Namew Lake gabbro (abandoned); PG, Pelican granulite (abandoned); PL, Peter Lake gabbro (Symons, 1994); RE, Reynard Lake granodiorite (Symons, 1995); SG, Sahli granulite (Gala et al., 1998); TP, Thompson pegmatites (measured); WB, Wathaman Batholith (Symons, 1991); WD, Wapisu Lake dome (Symons and Harris, 2000). b) Locations of collections near Thompson, Manitoba.

- There may be a 'hairpin' or approximately 90° change in direction of the APWP following this stillstand (Fig. GS-7-2; Symons, 1998; Symons and MacKay, 1999; Symons and Harris, 2000).

These results are predominantly for the THO because the Superior craton is represented paleomagnetically by only two paleopoles, at 1880 and 1849 Ma (Fig. GS-7-2). Therefore, the following four rock units were sampled from the Superior Boundary Zone in an attempt to better constrain the tectonic motions for the Superior craton between ca. 1840 and 1770 Ma: the ca. 1836 Ma Mystery Lake pluton, the ca. 1822 Ma Wintering Lake pluton, the ca. 1818 Ma Fox Lake pluton, and some ca. 1770 Ma pegmatites and associated contact zones (Harris et al., 1998, 1999a).

The Mystery Lake pluton has yielded several interesting paleomagnetic results (Fig. GS-7-3; Harris et al., 1999b). First, sites from the northern two-thirds of the pluton (mean (N)) yielded a paleopole suggesting that this portion of the pluton is older than 1836 Ma. Second, the first result is supported by the results from a northeast-trending mafic dyke at site 06 that yielded a paleopole location similar to the b-pole published for the 1880 Ma Molson dyke swarm (Halls and Heaman, 2000). Third, the sites from the southern portion of the pluton (mean (S)) yielded a paleopole direction more akin to the 1830–1810 Ma stillstand and appear to date a metamorphic event. Finally, a northwest-trending dyke at site 12 yielded a remanence direction similar to the ca. 1270 Ma Mackenzie dyke swarm direction that also has a northwest trend in the study area (Fig. GS-7-3). A geochronological study is underway on the Mystery Lake pluton and surrounding supracrustal rocks (Peck, 1999).

This report details the petrography, paleomagnetism and geobarometry of rocks within and around the Fox Lake pluton.

GEOLOGY

The Fox Lake pluton is located some 100 km northeast of Thompson (Fig. GS-7-1), within the Split Lake Block (SLB) on the northwestern margin of the Superior craton. The SLB is a dominantly

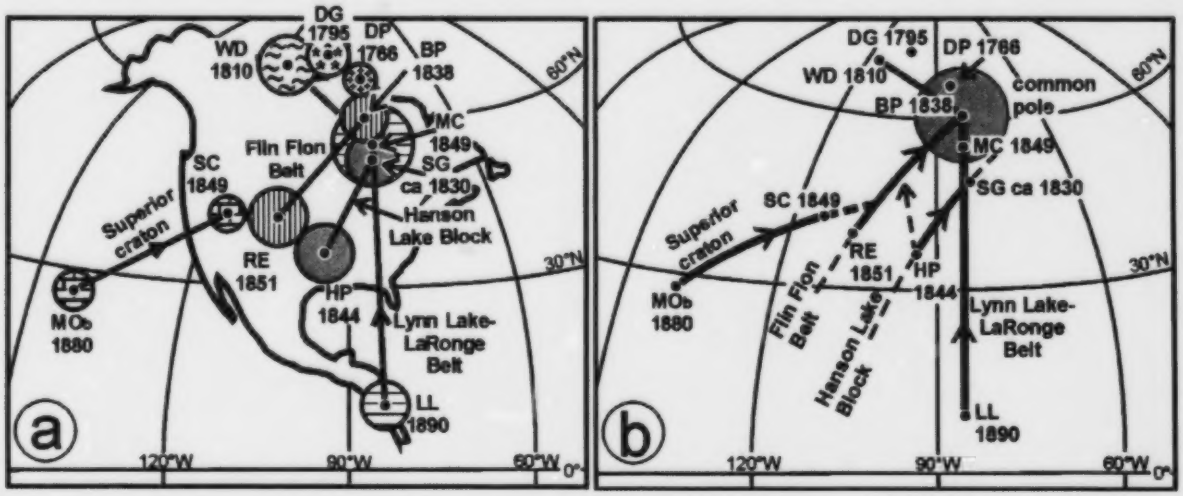


Figure GS-7-2: a) Apparent polar wander path (APWP) segments for domains of the Superior craton and the Trans-Hudson Orogen between ca. 1890 and 1766 Ma. Abbreviations for poles as noted in Figure GS-7-1, with their magnetization ages in Ma: MOb, Molson swarm dykes' b-pole (Halls and Heaman, 2000); SC, Sudbury Complex (Morris, 1984). b) Projection of the APWP segments to the ca. 1840 to 1810 stillstand and hairpin (Symons and Harris, 2000).

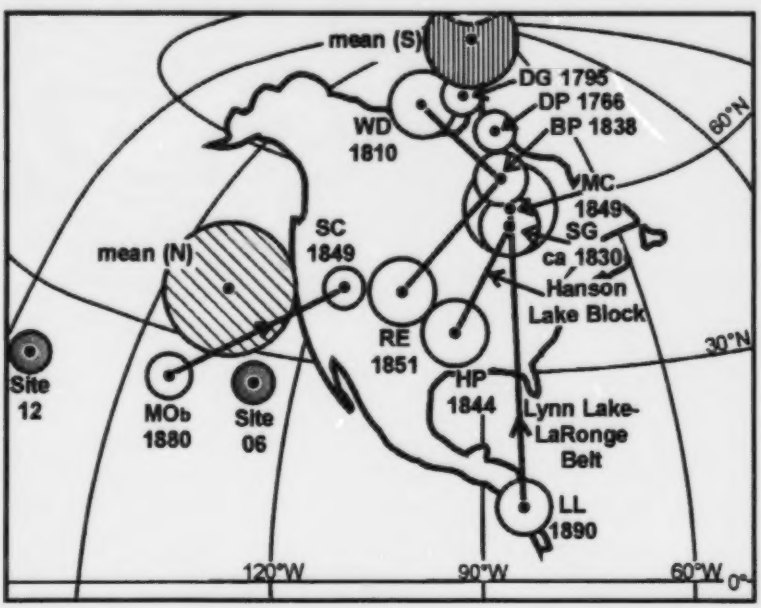


Figure GS-7-3: Apparent polar wander path (APWP) segments for domains of the Superior craton and Trans-Hudson Orogen, with the paleopole locations for the Mystery Lake pluton (after Harris et al., 1999b). Abbreviations as in Figure GS-7-1; mean (S), mean (N), site 06 and site 12 are referred to in the text.

granulite facies terrane that consists mainly of Archean meta-igneous rocks with traces of supracrustal rocks. The SLB contains evidence for pre-2.8 Ga protoliths, 2.71 Ga granodiorite magmatism and at least two late-Archean, high-grade metamorphic events at ca. 2695 and 2640 Ma (Heaman et al., 1999; Böhm et al., 1999 and references therein). The SLB is structurally bounded by two discrete linear belts of cataclastic mylonitic rocks, the east-trending Aiken River and the northeast-trending Assean Lake deformation zones to the south and north, respectively (Böhm et al., 2000).

Although the Fox Lake pluton is mapped over a large area (more than 400 km^2), outcrop is very sparse and restricted to lakeshores because of glacial overburden (Fig. GS-7-4). The pluton is mapped as a biotite granite that is well foliated along Awupak Narrows–Little Assean Lake and more massive along the shores of Split Lake (Corkery, 1985).

The pluton, to the east and south, appears to have intruded into only one type of host rock, a hornblende-biotite gneiss derived from tonalite, amphibolite and metagabbro (Manitoba Energy and Mines, 1991). This gneiss was sampled for radiometric dating near the southeast corner of the Fox Lake pluton, and its granodiorite paleosome yielded zircon $^{207}Pb/^{206}Pb$ ages between 2822 and 2617 Ma (Böhm et al., 2000). The northern boundary of the pluton is the Assean Lake Shear Zone, which is dominated by layered and laminated mylonite and cataclastic gneiss up to 500 m thick (Corkery, 1985).

The Fox Lake pluton has several radiometric age dates. The 1811 Ma $^{207}Pb/^{206}Pb$ model age from titanite is interpreted by Heaman and Corkery (1996) to be a minimum estimate for the pluton's emplacement. A more recent determination on titanite yielded an age of 1825 ± 6 Ma (L.M. Heaman, pers. comm., 2000). A K/Ar age of 1720 Ma (Lowden, 1961) was obtained from a large raft of migmatite found within the pluton, suggesting metamorphism at that time (Corkery, 1985).

PETROGRAPHY

The Fox Lake plutonic rocks range in colour from medium brown to medium red, pink and grey, and they do not exceed medium grain size. The pluton is massive along the shores of Split Lake and very foliated along the shores of Awupak Narrows (Fig. GS-7-4). The foliation is defined by mafic minerals, most commonly biotite. Along Awupak Narrows, in the Assean Lake Shear Zone, the foliation within the pluton trends approximately 040° and dips to the southeast at 70 to 90°. This shear zone and the Aiken River Shear Zone along the southern shore of Split Lake are known as D_4 structures, and are found only around these two lakes. The Assean Lake Shear Zone is further affected by D_5 north-trending faults in the Little Assean Lake area (Corkery, 1985). On the western shore of Split Lake, the weak foliation within the pluton generally strikes approximately 130° and is subvertical.

Thirty samples of the pluton were examined in thin section, and each was point counted with a minimum 100 points. All mineral abundances are reported as modal per cent. The pluton consists mainly of monzogranite, with plagioclase more abundant than K-feldspar, and lesser amounts of tonalite, granodiorite and monzonite.

The relatively undeformed plutonic samples from sites 07 to 12 along Split Lake are generally medium grained and dominated by plagioclase, quartz and K-feldspar. Although igneous crystal contacts are visible, recrystallization mortar textures are ubiquitous, with coarser phenocrysts of plagioclase, quartz, and microcline or microperthite surrounded by very fine grained recrystallized quartz. Biotite, in amounts up to 15%, is often anhedral, ragged, partially altered to chlorite and occasionally crenulated. Other minerals occurring in trace to minor amounts (i.e. <2%) include apatite, titanite, zircon and undifferentiated opaque minerals. Deformation textures are uncommon, essentially being restricted to undulose extinction in quartz crystals. Besides biotite, plagioclase is the only mineral showing alteration, commonly to clay minerals.

The samples from Awupak Narrows and Little Assean Lake are very deformed and typically mylonitic, with lenses of undeformed granitoid surrounded by very thin (<1 mm wide) high-strain zones. The high-strain zones are generally characterized by fine-grained biotite and quartz, or by quartz ribbons if biotite is not present. Deformation and recrystallization are greatest at sites 13 to 15, and decrease to the southwest and northeast as the shoreline encroaches farther into the pluton (Fig. GS-7-4). At the more deformed sites, grain size is very fine, with few crystals exceeding 0.75 mm, and high-strain zones are predominant. Farther inward, at sites 06 and 20, grain size is variable, being finer along the high-strain zones and coarser between them, often up to 4 mm

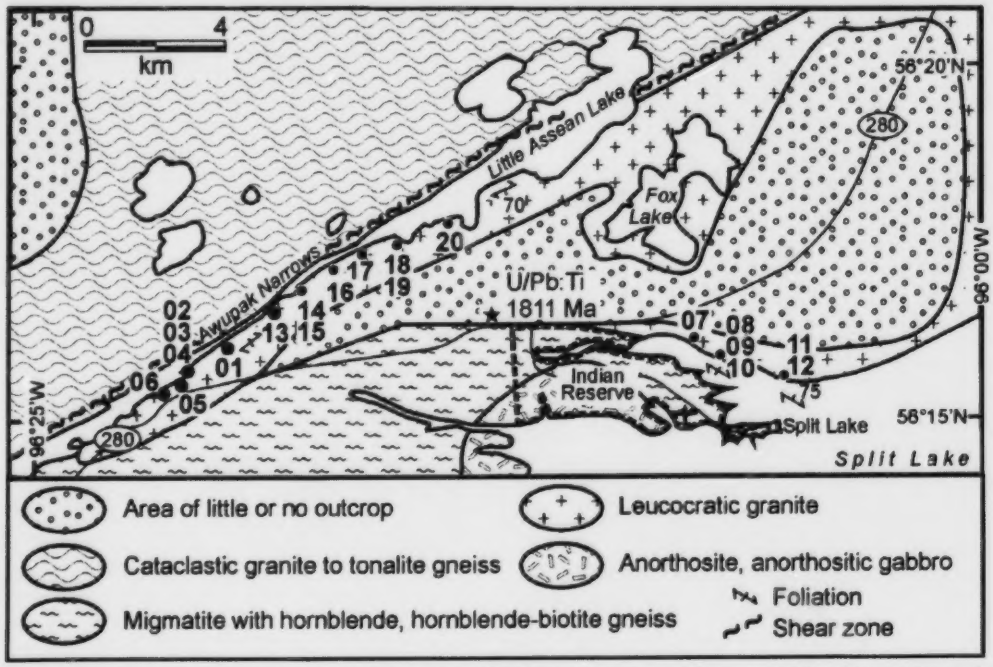


Figure GS-7-4: Local geology around the Fox Lake pluton (from Manitoba Energy and Mines, 1991) with locations of sampling sites. U/Pb:Ti is the radiometric model age from titanite (Heaman and Corkery, 1996).

in length. Quartz is generally the dominant mineral (30–60%), likely due to greater recrystallization in these samples compared to the Split Lake samples. Plagioclase rarely shows twinning and ranges from 15 to 40%, whereas microcline, orthoclase and perthite are present, in combination, up to 30%. Muscovite is present in all samples, likely as a product of feldspar hydrolysis during deformation (O'Hara, 1988; O'Hara and Blackburn, 1989). A few samples contain fine-grained rounded garnets indicative of the metamorphic conditions prevalent during deformation. Biotite never exceeds 10% when it is present, and is less common than muscovite. Other trace and minor minerals include zircon, apatite, titanite and undifferentiated opaque minerals. Plagioclase is commonly altered to clay minerals and biotite is often altered to chlorite.

The granitic gneiss host rock was sampled at sites 08 to 10 (Fig. GS-7-4). It is white-grey to cream-grey, medium to coarse grained and generally foliated, the foliations being defined by biotite. Plagioclase is generally more common than quartz, microcline and microperthite. Biotite is present (5–20%) and there are trace amounts of apatite, zircon, titanite and undifferentiated opaque minerals.

The amphibolite collected at site 02 is likely a predeformational mafic dyke into the Fox Lake pluton. It is fine grained, dark green and foliated, although the foliation is defined by the intervening matrix around randomly oriented anhedral amphiboles. Minor amounts of quartz, plagioclase, epidote, clay minerals and undifferentiated opaque minerals are also present.

The mafic dyke collected at site 09 crosscuts the granitic gneiss and shows no deformation. This amphibolite is fine to medium grained and dark green-grey, with a moderately well developed lineation defined by amphibole. The rock consists of more than 80% amphibole, with subordinate amounts of plagioclase and quartz, and traces of epidote and undifferentiated opaque minerals.

The deformed mafic dyke collected at site 10 is older than the Fox Lake pluton but younger than the D_3 event that likely metamorphosed the hosting granitic gneiss. This amphibolite is medium grained and dark green-grey, with a moderately well developed lineation that is defined by amphibole grains. Plagioclase and quartz are relatively equal in proportion, about 15% each, and somewhat recrystallized. Traces of undifferentiated opaque minerals are also present.

At site 12, the Fox Lake pluton has an intrusive contact with deformed, amphibolitic gneiss host rocks. The gneiss is medium grained and dark green-grey, with a well developed foliation defined by amphibole and biotite. The microtexture comprises medium-grained (1–2.5 mm) amphibole (50%) with associated biotite (<15%), separated from other amphiboles by fine-grained recrystallized plagioclase (<15%) and quartz (<10%). Very fine grained epidote is abundant and generally associated with the felsic matrix.

PALEOMAGNETISM

Methodology

The Fox Lake pluton was sampled at 20 sites (Fig. GS-7-4) for paleomagnetic measurements, 15 of which sampled only the pluton. Two sites sampled both the granitoid and the host granitic gneiss (sites 08 and 10). The remaining three sites sampled the granite with deformed amphibolite (site 02), a mafic dyke with the host granitic gneiss (site 09), and the granite with an amphibolite host rock (site 12). The five sites that sampled multiple rock types were collected as paleomagnetic contact tests in an attempt to obtain information on the relative timing of magnetization acquisition.

At each site, four to ten (average six) standard 2.5 cm diameter cores were drilled and oriented in situ, mostly using a solar compass, with periodic checks using a magnetic compass. The cores were later sliced into 2.2 cm lengths, yielding 173 specimens for paleomagnetic measurements. The specimens were stored in a magnetically shielded room for four months to allow their unstable viscous magnetizations to decay.

All paleomagnetic measurements were made in the Paleomagnetic Laboratory at the University of Windsor. All instruments used are housed in a three-layer magnetically shielded room that has an ambient magnetic field of about 0.2% that of the Earth's field intensity. Analyses of the data used the end-point method of Kirschvink (1980) to calculate the specimens' characteristic remanent magnetization (ChRM) directions. The ChRM for each specimen was accepted for statistical analysis if the best-fit stable direction (*see* Fig. GS-7-7) had a maximum angular deviation of less than 18° over three or more consecutive steps, although most had less than 10° deviation. Site and pluton mean directions were calculated using the statistical methods of Fisher (1953).

First, the specimens' natural remanent magnetization (NRM) was measured using a Canadian Thin Films (CTF) DRM-420 fully automated cryogenic magnetometer, yielding a median magnetic intensity for the granitoid rocks of 1.5×10^{-3} amperes/metre (A/m; first quartile, $Q_1 = 7.1 \times 10^{-4}$ A/m; third quartile, $Q_3 = 4.2 \times 10^{-3}$ A/m). Two specimens from each rock type per site were then selected as pilot specimens, each having 'average' NRM direction and intensity for the rock type and site. One pilot specimen underwent detailed demagnetization in an alternating field (AF), in 11 steps from 5 to 130 milliTesla (mT), using a Sapphire Instruments SI-4 AF demagnetizer. The other pilot specimen underwent detailed thermal demagnetization, in 11 steps from 200° to 570°C, using a Magnetic Measurements MMTD-80 thermal demagnetizer.

Thermal decay curves for the pilot specimens suggested that the magnetic carriers may: 1) be titanomagnetite with a gradual intensity decay over the entire temperature range (Fig. GS-7-5, specimens 04, 05, 17, 08Gr); 2) be pyrrhotite with large intensity decays between 200° and 320°C (Fig. GS-7-5, specimens 08hG, 12hA); and, 3) possibly have a hematite component with the titanomagnetite (Fig. GS-7-5, specimen 07). Comparison between the two pilot specimens per site suggested that all remaining specimens should be equally split between AF and thermal demagnetization techniques, using schedules similar to those used for the pilot specimens.

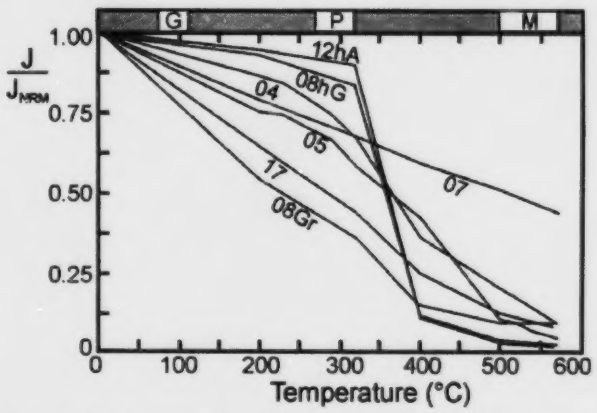


Figure GS-7-5: Thermal-decay curves for representative specimens of the Fox Lake area. Numbers refer to the sampling site: Gr, Fox Lake granitoid; hA, host amphibolite; hG, host granitic gneiss. Typical unblocking temperature ranges shown for goethite, pyrrhotite and magnetite (G, P and M, respectively). Vertical axis, J/J_{NRM} is the ratio of the measured intensity to its NRM intensity.

Sixteen of the 27 AF pilot specimens were further tested for their magnetic mineralogy and domain size using saturation isothermal remanent magnetization (SIRM) techniques. These specimens were magnetized in a direct field, in 14 steps from 10 to 900 mT, using a Sapphire Instruments SI-6 pulse magnetizer, and then AF demagnetized in 8 steps from 10 to 140 mT. Acquisition and decay curves from the SIRM technique support the presence of magnetite that is single domain in nature (Fig. GS-7-6, specimens 08hG, 10Gr, 10mD). However, SIRM testing

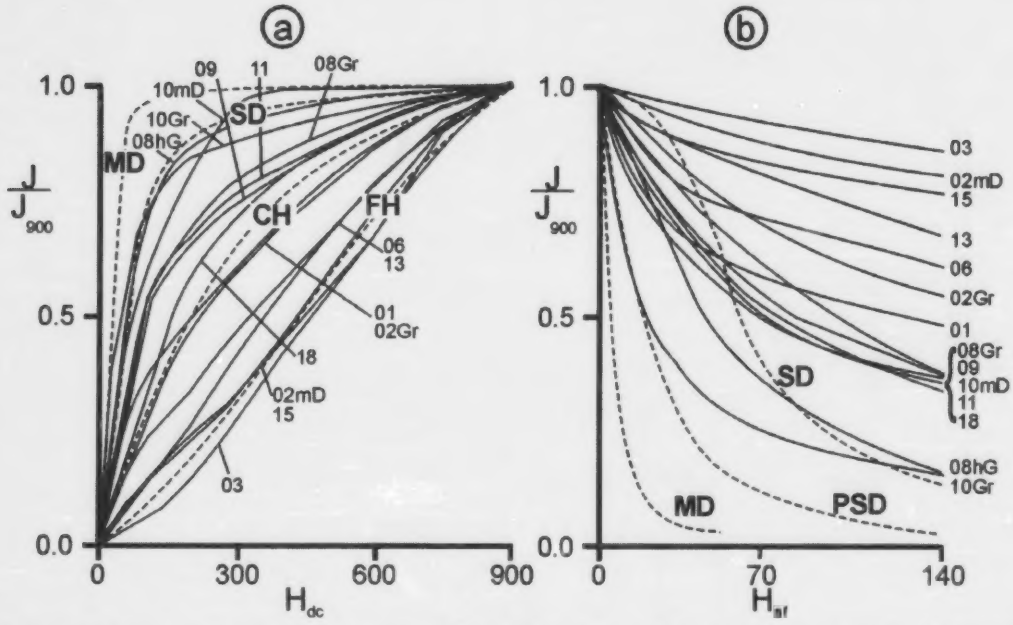


Figure GS-7-6: Saturation isothermal remanent magnetization (SIRM) acquisition (a) and decay (b) curves for the Fox Lake area. H_{dc} and H_{af} are the applied magnetizing and demagnetizing fields, respectively. Vertical axis, J/J_{900}, is the ratio of measured remanence intensity to isothermal saturation intensity at 900 mT. Type curves (dashed) are for multidomain (MD), pseudo–single domain (PSD) and single domain (SD) magnetite, and for coarse-grained (CH) and fine-grained (FH) hematite. Abbreviations: mD, mafic dyke; other abbreviations as in Figure GS-7-5.

indicates a more significant presence of hematite than shown by the thermal decay curves. Further, it appears the hematite has a complete magnetic crystallinity range from fine to coarse grained (Fig. GS-7-6).

About 70 granitoid specimens had their magnetic susceptibility measured using a Sapphire Instruments SI-2b Magnetic Susceptibility and Anisotropy Meter. They yielded a range of values between 0 and 35 000 SI volume units (dimensionless); however, 85% of the specimens yielded values of less than 150 SI volume units. This weak susceptibility is similar to that found for the granitoid specimens of the Mystery Lake pluton (Harris et al., 1999b).

Results

Typical direction and intensity changes for several specimens are shown in orthogonal step demagnetization plots (Fig. GS-7-7; Zijderveld, 1967). Any viscous remanent magnetization from recent overprinting that still remained after sitting in the field-free space was quickly removed within the first couple of demagnetization steps. Following the removal of any residual viscous component, the specimens' remanences generally decayed linearly to the origin on the vector plots, irrespective of demagnetization technique or rock type (Fig. GS-7-7).

Site mean ChRM directions were calculated for all sites (Table GS-7-1). Those with cones of 95% confidence (α_{95}) of greater than 20° are deemed to be statistically weak and are not considered for further discussion. All but one site yielded normal-polarity (positive, down) mean ChRM directions and only about 15% of all the specimens measured yielded reversed polarity (negative, up) ChRM directions. The majority of the sites have steep inclinations with a variety of declinations (Fig. GS-7-8). The direction of the reversed site (Table GS-7-1, site 09ii) is not antipodal to the normal site-mean directions and is clearly significantly different from the majority of other directions; it is therefore not considered further. Site 09i, from the host granitic gneiss, also gave a statistically different direction from the mean of the remaining normal-polarity sites, so it too is not considered further.

Contact Tests

Five sites were collected to provide contact tests to help determine the relative timing in acquisition of magnetization. Site 02 sampled the granitoid and a deformed mafic dyke. The test failed, however, because both rock units yielded statistically similar ChRM directions (Table GS-7-1; Fig. GS-7-8). This was the only test site found on Awupak Narrows–Little Assean Lake.

Sites 08, 09 and 10 are all in a similar location, separated by less than 100 m east to west, but the outcrops are discontinuous. The contact test at site 08 was between the granitic gneiss and the pluton. The two ChRM mean directions are statistically different, which suggests that the two rock types were magnetized at different times. Site 09 sampled the granitic gneiss and an undeformed mafic dyke. This contact test is deemed to be inconclusive. Both rock types produced bimodal directional groups. The mafic dyke had two cores yielding directions to the northwest and inclined moderately upward, and three cores with directions to the east and west and steeply upward and downward, respectively. The granitic gneiss had three specimens that yielded northerly and moderately downward directions and three other specimens with directions to the southeast and steeply downward. Site 10 sampled the granitic gneiss, an early weakly deformed mafic dyke, and the pluton. All three gave statistically similar directions (Fig. GS-7-8), providing a negative contact test and suggesting a similar age for their acquisition of magnetization.

Site 12 sampled the pluton and deformed amphibolitic gneiss host rock. Disregarding one aberrant direction from a monzonite specimen, the two rock types gave statistically similar ChRM directions, suggesting similar magnetization ages.

GEOBAROMETRY

The Fox Lake pluton was sampled to estimate pressure-temperature conditions, and thus the amount of uplift and potential tilt of the pluton during uplift. The latter permits interpretation of the paleomagnetic data with greater confidence. Bedding planes are usually used to determine

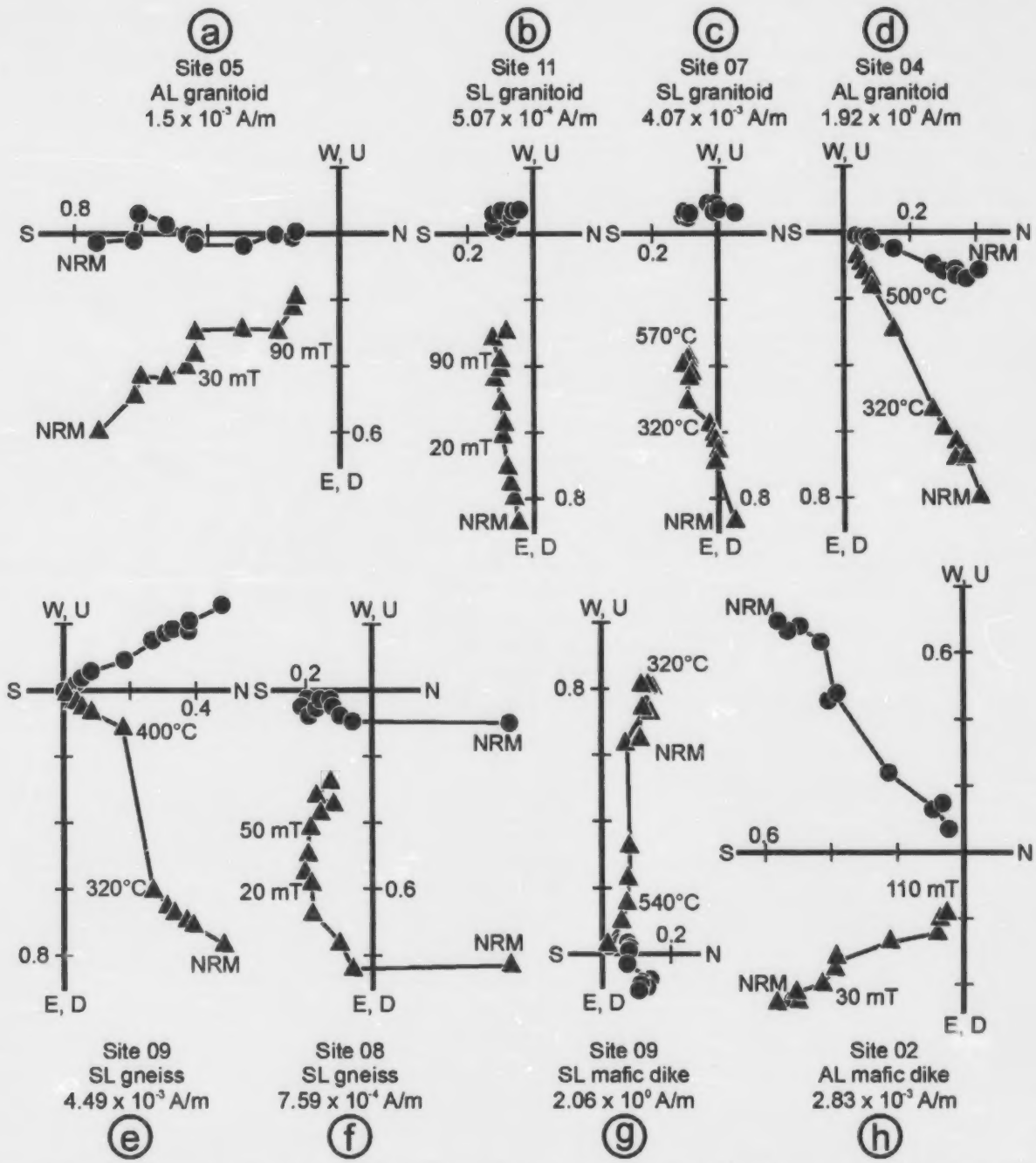


Figure GS-7-7: Representative orthogonal step demagnetization plots (Zijderveld, 1967) for selected specimens from the Fox Lake study. Circles are projected in the horizontal plane (east-south-west-north) and the triangles in the vertical plane (down-south-up-north). Site number, location (AL, Assean Lake; SL, Split Lake), rock type and NRM intensity are given.

paleohorizontal for supracrustal rocks, but this is rarely possible with plutonic rocks. However, geobarometric techniques can be used to calculate the pressures of crystallization or recrystallization for a particular mineral assemblage in the pluton. The horizontal pressure gradient across the pluton is then obtained, from which any postcrystallization tilting can be estimated.

The Al-in-hornblende geobarometer was the intended tool because the required mineral and chemical systems for it are generally available in orogenic granitic rocks. This geobarometer was initially calibrated experimentally and empirically for calc-alkaline granitoid rocks containing the mineral assemblage quartz+alkali feldspar+plagioclase +hornblende+biotite+Fe-Ti oxide+titanite+melt+fluid (Johnson and Rutherford, 1989; Thomas and Ernst, 1990; Schmidt, 1992). However, the effect on the pressure calculation of the absence of alkali feldspar and titanite has been questioned (Anderson and Smith, 1995). The possibility of not needing alkali feldspar in the mineral assemblage makes this geobarometer more useful by expanding the number of rock types for which it may work.

Table GS-7-1: Site-averaged paleomagnetic data for the Fox Lake pluton.

Site	Rock type[a]	Spec. N/n[b]	Dec.[c] (°)	Inc. (°)	α_{95} (°)	k	Statu	Location
01	Gr	4, 3	159.8	77.9	28.2	20.1	Reject	Assean Lake
02	mD	8, 6	344.2	69.9	12.7	29.0	Accept	Assean Lake
	Gr	4, 3	314.5	73.1	14.0	79.0	Accept	
03	Gr	8, 6	267.9	60.3	13.1	19.0	Accept	Assean Lake
04	Gr	8, 5	040.4	63.8	19.9	15.7	Accept	Assean Lake
05	Gr	6, 4	047.8	72.4	24.1	15.5	Reject	Assean Lake
06	Gr	9, 6	109.8	83.6	26.4	7.4	Reject	Assean Lake
07	Gr	9, 9	208.9	67.2	6.1	72.2	Accept	Split Lake
08	hG	5, 5	157.0	1.8	10.1	58.0	Accept	Split Lake
	Gr	6, 6	025.7	89.4	5.6	144.2	Accept	
09i	mD	5, 5	358.1	85.6	20.3	15.1	Accept	Split Lake
ii	mD	3, 3	325.5	-17.6	19.1	42.6	Reject	
i	hG	3, 3	347.9	53.9	17.7	49.6	Reject	
ii	hG	3, 3	150.0	76.2	7.9	244.3	Accept	
10	Gr	7, 6	145.4	72.4	15.7	19.3	Accept	Split Lake
	hG	2, 2	172.2	62.4	14.2	310.8	Accept	
	mD	6, 6	167.1	76.4	9.8	47.9	Accept	
11	Gr	11, 10	150.7	72.4	10.6	21.7	Accept	Split Lake
12	Gr	4, 3	221.9	73.6	19.5	41.2	Accept	Split Lake
	hA	8, 8	176.7	72.4	6.3	79.1	Accept	
13	Gr	6, 4	339.3	76.2	15.7	35.3	Accept	Assean Lake
14	Gr	6, 6	011.9	55.1	51.7	2.6	Reject	Assean Lake
15	Gr	6, 4	249.5	34.7	26.0	13.4	Reject	Assean Lake
16	Gr	6, 6	280.1	51.4	53.4	2.5	Reject	Assean Lake
17	Gr	5, 4	302.7	82.6	11.7	62.1	Accept	Assean Lake
18	Gr	9, 9	121.5	84.2	17.8	9.3	Accept	Assean Lake
19	Gr	7, 7	044.2	83.4	6.2	96.0	Accept	Assean Lake
20	Gr	7, 7	182.1	85.5	11.9	26.9	Accept	Assean Lake

[a] Gr, granite; mD, mafic dike; hA, host amphibolite; hG, host granitic gneiss

[b] 'Spec. N/n' refers to the number of specimens measured (N) and the number of specimens used in the mean ChRM direction

[c] Mean directions are given by their declination (Dec.), inclination (Inc.), radius of 95% cone of confidence (α95), and precision parameter (k), based on the statistical methods of Fisher (1953)

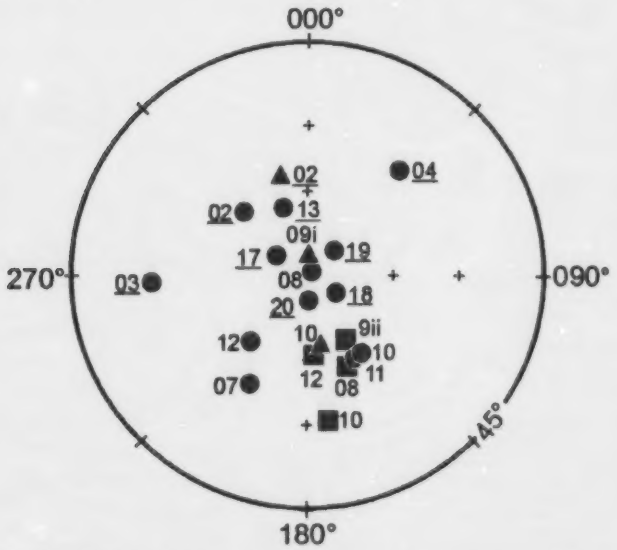


Figure GS-7-8: Central part of an equal-area stereogram showing the site mean ChRM directions, with site numbers labelled. Underlined numbers are sites from Awupak Narrows and nonunderlined numbers are sites from Split Lake. Circles are means from granitoid sites, triangles are means from mafic dykes, and squares are means from host rocks (Table GS-7-1).

This latter point is very fortunate because, although the Fox Lake pluton does not contain any hornblende, the mafic rocks do contain abundant hornblende. Also, conveniently, the mafic rocks at sites 09 and 12 contain the remainder of the mineral assemblage and were therefore analyzed for amphibole-plagioclase pairs.

Methodology

Chemical analyses on hornblende and plagioclase pairs were done on the JEOL JXA-8600 Superprobe at the University of Western Ontario. Operating conditions were set at a 15 kV accelerating voltage, a 10 nA beam current and a spectrum accumulation time of 20 s. Electron beam focus diameter was 2 µm for amphibole and 5 µm for plagioclase (*see* Harris et al., 1997, 1999b for complete procedures). Three amphibole-plagioclase crystal pairs were analyzed on one polished thin section per site.

Following chemical analysis, structural formulas were calculated for the amphibole and plagioclase. A preliminary pressure was calculated using the total aluminum in the amphibole and the calibration of Schmidt (1992). The temperature was then determined using the preliminary pressure estimate and the plagioclase-amphibole geothermometer of Blundy and Holland (1990). Finally, the temperature and total aluminum of the amphibole were used to determine the final pressure, according to the temperature-corrected geobarometer of Anderson and Smith (1995).

Results

Representative compositions of amphibole and plagioclase are given in Tables GS-7-2 and -3. Structural formulas for the amphibole were calculated on the basis of 13 cations exclusive of Ca, Na and K (Cosca et al., 1991; Harris et al., 1997) and, for the plagioclase, on the basis of 32 oxygen atoms. The amphibole from site 09 is generally not alkali-rich (i.e. $(Na + K)_A < 0.50$) and is classified as magnesiohornblende and tschermakite (Fig. GS-7-9). The amphibole from site 12 is much more alkali-rich (i.e. $(Na + K)_A \geq 0.50$) and is classified as edenite or magnesiohastingsite. Plagioclase ranges from An_{26} to An_{37}. The $Fe^{3+}/[Fe^{3+}+Fe^{2+}]$ ratio ranges from 0.15 to 0.28 and the $[Fe^{3+}+Fe^{2+}]/[Fe^{3+}+Fe^{2+}+Mg]$ ratio ranges from 0.49 to 0.53. Preliminary pressures range from 660 to 761 MPa, and temperatures range from 745 to 783°C (Table GS-7-4). Final temperature-corrected pressures range from 533 to 592 MPa and the corresponding depths have a very narrow range of 19.3 to 21.3 km (Table GS-7-4).

DISCUSSION

The crystallization age of the Fox Lake pluton is not well constrained. The U/Pb titanite model ages of 1811 and 1825 Ma are the minimum ages of emplacement (Heaman and Corkery, 1996; L.M. Heaman, pers. comm., 2000); the true age may therefore be somewhat older. We consider the average, 1818 Ma, as the minimum emplacement age and potential age of magnetization acquisition. The only other constraint on the age of the pluton is that it must be older than the cataclastic deformation caused by the Assean Lake Shear Zone (D_4), which is, in turn, only constrained in age by the Fox Lake pluton. It is noted that mafic dykes corresponding to the 1880 Ma Molson dyke swarm (Halls and Heaman, 2000) are mapped within the southern margin of the pluton and within the Assean Lake Shear Zone (Corkery, 1985; Manitoba Energy and Mines, 1991). If these are truly dykes of the Molson swarm, then the pluton is older than 1880 Ma; however, there is geological evidence that the pluton does, in fact, cut the Molson dykes (M.T. Corkery, pers. comm., 2000).

Accepting that the Fox Lake pluton predates the D_4 event and is not syndeformational, the statistical F-test of McFadden and Lowes (1981) was used to compare the site mean directions of the pluton from the Awupak Narrows–Little Assean Lake area to those from the Split Lake

Table GS-7-2: Representative amphibole analyses for mafic rocks from the eastern zone of the Fox Lake pluton.

Site:	FX09	FX09	FX12	FX12
Analysis #:	2.2	3.2	2.2	3.2
SiO_2 (wt.%)	41.97	42.00	41.69	42.50
TiO_2	0.60	0.59	0.61	0.49
Al_2O_3	11.66	11.36	11.74	11.69
Fe_2O_3*	0.00	0.00	0.00	0.00
Cr_2O_3	0.07	0.02	0.03	0.02
FeO	17.97	17.44	16.98	17.06
MnO	0.28	0.21	0.23	0.30
MgO	9.08	9.31	9.72	9.88
CaO	11.89	11.55	11.84	12.02
Na_2O	1.16	1.16	1.34	1.25
K_2O	1.01	0.99	1.12	1.09
F	0.00	0.00	0.03	0.00
Cl	0.03	0.02	0.00	0.09
Total	95.72	94.65	95.33	96.39
Si (ions)	6.417	6.462	6.384	6.431
Al	1.583	1.538	1.616	1.569
Total Z	8.000	8.000	8.000	8.000
Al	0.519	0.523	0.504	0.516
Ti	0.069	0.068	0.070	0.056
Cr	0.008	0.003	0.003	0.003
Fe^{3+}	0.480	0.528	0.468	0.464
Fe^{2+}	1.818	1.716	1.707	1.694
Mg	2.069	2.135	2.218	2.228
Mn	0.037	0.028	0.030	0.039
Total M1,2	5.000	5.000	5.000	5.000
Ca	1.948	1.904	1.943	1.949
Na	0.052	0.096	0.057	0.051
Total M4	2.000	2.000	2.000	2.000
Na	0.293	0.250	0.339	0.316
K	0.197	0.194	0.218	0.210
Total A	0.490	0.444	0.558	0.525
OH*	1.993	1.994	1.983	1.977
F	0.000	0.000	0.017	0.000
Cl	0.007	0.006	0.000	0.023
Fe^{3+}#	0.53	0.51	0.50	0.49
Fe#	0.23	0.25	0.23	0.23

Fe_2O_3* and OH* are back-calculated from the structural formula

Fe^{3+}# = $Fe^{3+}/(Fe^{3+} + Fe^{2+})$

Fe# = $(Fe^{3+} + Fe^{2+})/(Fe^{3+} + Fe^{2+} + Mg)$

Analysis # comprises crystal number and spot number

Table GS-7-3: Representative plagioclase analyses for mafic rocks of the Fox Lake area.

Site:	FX09	FX09	FX12	FX12
Analysis #:	2.2	3.2	2.2	3.2
SiO_2 (wt.%)	58.61	58.15	58.44	59.38
Al_2O_3	25.18	25.46	25.16	24.82
Fe_2O_3	0.29	0.38	0.30	0.25
CaO	7.30	7.40	7.05	6.42
Na_2O	7.39	6.68	7.28	7.55
K_2O	0.16	0.15	0.06	0.17
Total	98.94	98.22	98.29	98.59
Si (ions)	10.587	10.559	10.604	10.721
Al	5.361	5.449	5.381	5.282
Fe^{3+}	0.040	0.052	0.041	0.035
Total Z	15.987	16.059	16.026	16.038
Ca	1.413	1.440	1.371	1.242
Na	2.588	2.351	2.561	2.643
K	0.038	0.035	0.015	0.038
Total A	4.038	3.826	3.946	3.923
An	34.99	37.63	34.73	31.66
Ab	64.08	61.46	64.89	67.37
Or	0.93	0.90	0.37	0.97

Analysis # comprises crystal number and spot number

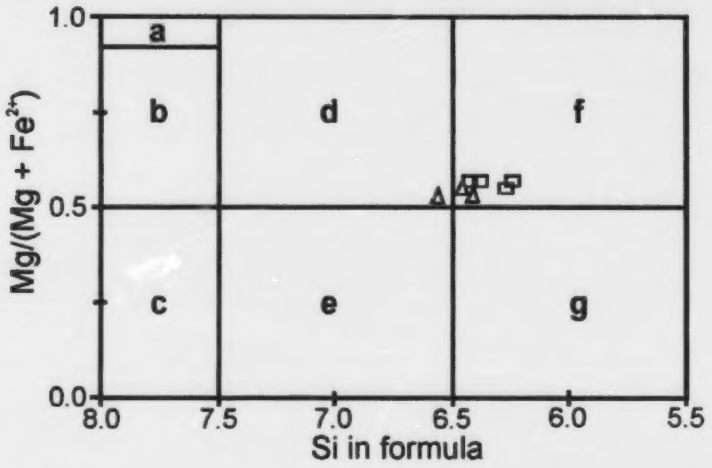
area, bearing in mind the latter specimens were much less deformed. The test proved negative, showing that the mean directions for the two populations of granitoids were statistically different at the 95% confidence level (i.e. [F_{calc} = 4.47] > [$F_{tabulated}$ = 3.98]; Table GS-7-5, sets 1 and 3). Therefore, it is concluded that two magnetizations are present, one acquired during the D_4 event and one earlier, likely during crystallization of the Fox Lake pluton. Further, recalling that most of the contact tests failed, the other host rock and dyke types also carry directions similar to those of the pluton in two areas, along the Assean Lake Shear Zone and along the Split Lake shoreline (Fig. GS-7-4, -8; Table GS-7-1; Table GS-7-5, sets 2 and 4). Again, when all sites are used to compare the mean directions between the two areas, they give statistically different directions (i.e. [F_{calc} = 11.75] > [$F_{tabulated}$ = 3.55]). This results in four mean directions to be considered.

The paleopole locations for the mean directions were calculated (Table GS-7-5) and plotted (Fig. GS-7-10). Note that, for each of the Awupak Narrows and Split Lake areas, the granitic rocks alone and all sites combined have similar paleopole directions (Fig. GS-7-10a, sets 1-2 and 3-4), indicating similar magnetization ages within each area. The first granite pole, corresponding to the Awupak Narrows–Little Assean Lake area, is close to the pole calculated for the Wapisu gneiss dome at 1810 Ma (Fig. GS-7-10b; Symons and Harris, 2000) and is in agreement with the average titanite age of 1818 Ma for the Fox Lake pluton. This implies that the paleopole and model age records recrystallization during metamorphism and deformation in the Assean Lake Shear Zone.

The location of the other granite pole from the Split Lake sites is similar to three other THO paleopoles with ages ranging from 1851 to 1844 Ma (Fig. GS-7-2, -10b). This implies that this pole likely records magnetization during primary crystallization at ca. 1850 Ma.

There are three hypotheses that could explain why the two ca. 1850 Ma paleopoles for the Superior craton, from the Sudbury Complex and the Fox Lake pluton, are close but not at precisely the same location. First, the paleopole for the Split Lake sites of the Fox Lake pluton is derived from a small sample set of five sites, and therefore may not be truly representative. Second, the paleopole calculated for the Sudbury Complex (Morris, 1984) is confidently dated at 1849 Ma (Krogh et al., 1984; Corfu and Lightfoot, 1996; Ames et al., 1998); however, it required structural corrections to account for various rotations around the basin. Any potential errors in measuring and correcting for the rotations will lead to minor changes in the location of the paleopole. Third, post-1850 Ma rotations associated with tectonism between the two collection localities would also lead to discordance between the two poles. Examples of such rotations could include 1) relative rotation of the Fox Lake area during final closure of the THO; 2) rotation between Fox Lake and Sudbury during the ca. 1100 Ma Keweenawan rifting event; or 3) rotation of the Sudbury Complex as a result of final collision during the Grenville orogeny. Therefore, there is a strong implication that the Fox Lake pluton was emplaced ca. 1850 Ma, but additional radiometric or other age constraints are certainly needed.

The geobarometry data are somewhat speculative because of the use of the Al-in-hornblende geobarometer on mafic rocks with mineral assemblages for which it has not been specifically calibrated. However, this is the first P-T data for the area, a deficiency noted by Böhm et al. (1999). These values should be considered with caution because K-feldspar and titanite were missing from required mineral assemblage for the Al-in-hornblende geobarometer, although quartz+plagioclase+amphibole+biotite+apatite+Fe-Ti oxide were present. Still, the anorthite content (An) of the plagioclase, the trivalent iron ratio [Fe^{3+}#] and the total iron ratio [Fe#] are all within the ranges suggested by Anderson and Smith (1995) and Anderson (1996) to ensure that the correct temperature



Ca_{M4} > 1.50 - calcic amphibole		
	$(Na + K)_A$ < 0.50 Δ	$(Na + K)_A$ > 0.50 □
a	tremolite	———
b	actinolite	———
c	ferro-actinolite	———
d	magnesiohornblende	edenite
e	ferrohornblende	ferro-edenite
f	tschermakite	magnesiohastingsite ($^{VI}Al < Fe^{3+}$) pargasite ($^{VI}Al > Fe^{3+}$)
g	ferrotschermakite	hastingsite ($^{VI}Al < Fe^{3+}$) ferropargasite ($^{VI}Al > Fe^{3+}$)

Figure GS-7-9: Part of the calcic-amphibole classification scheme (Leake et al., 1997) and data from sites 09 (triangles) and 12 (squares).

Table GS-7-4: Summary of site-averaged compositions and standard deviations of coexisting amphibole and plagioclase edges and thermobarometric data for the Fox Lake area.

Site	FX09	FX12
N	3	4
Plagioclase (An #)[a]	34.6 ± 1.0	32.1 ± 2.1
Amphibole (average Al_{TOT})	2.05 ± 0.02	2.18 ± 0.06
P (MPa)[b]	674 ± 10	736 ± 28
T (°C)[c]	753 ± 5	768 ± 11
P (MPa)[d]	536 ± 3	555 ± 22
Depth (km)	19.3 ± 0.1	20.0 ± 0.8

[a] An/(Ab+An+Or)
[b] Schmidt (1992)
[c] Blundy and Holland (1990)
[d] Anderson and Smith (1995)

would be calculated and that oxygen fugacity would not become an additional factor. Ultimately, further experimentation on the Al-in-hornblende geobarometer for other non–calc-alkaline granitoid rock types will establish the reliability of these results.

With these thoughts in mind, a few observations are worthwhile. The two mafic samples are not necessarily of similar age. FX09 is from an undeformed mafic dyke that clearly cuts the granitic gneiss, but the contact relation with the Fox Lake pluton is unknown. FX12 is a deformed amphibolite that is intruded by the pluton and considered Kenoran (ca. 2695 Ma) in age (M.T. Corkery, pers. comm., 2000). However, both sites give similar calculated temperatures of 753° ± 5° and 768° ± 11°, and pressures of 536 ± 3 MPa and 555 ± 22 MPa. Therefore, they likely record metamorphic conditions corresponding to medium-pressure amphibolite grade. These pressures and temperatures measured for the mafic samples represent conditions similar to those found in other parts of the Superior Boundary Zone (e.g. the Thompson Block, at 700–750°C, 600–700 MPa, and the Pipe Lake Block, at 600–660°C, 500–600 MPa; Weber, 1990). Finally, the statistically identical depth estimates of 19.3 ± 0.1 km and 20.0 ± 0.8 km suggest that no tilt corrections should be applied to the paleomagnetic directions before calculating their paleopoles. Note that this metamorphic event most likely predates intrusion of the Fox Lake pluton because: 1) approximately 20 km is rather deep for felsic intrusions (Hess, 1989); and 2) the last high-grade metamorphic event is suggested to have occurred around 2695 Ma, which was followed by greenschist-grade metamorphic overprinting around 1825 Ma (Böhm et al., 1999).

Finally, the metamorphic mineralogy of the mylonitized rocks in

Table GS-7-5: Summary of paleomagnetic data for the Fox Lake pluton and associated rocks.

Set	N	Dec,Inc (°)	α95 (°)	R	PLAT,PLON (°N) (°W)	dp,dm (°)	Notes
1	8	330, 83	11.3	7.722	67, 113	22,22	AL, Gr, sites 2-4, 13, 17-20
2	9	334, 82	10.2	8.696	69, 116	19,20	AL, all, sites 2-4, 13, 17-20
3	5	183, 78	12.1	4.902	33, 97	21,23	SL, Gr, sites 7, 8, 10-12
4	11	173, 77	6.4	10.806	31, 87	18,18	SL, all, sites 7-12

N, number of sites in mean
Dec, declination; Inc, inclination
R, resultant vector
PLAT, paleolatitude; PLON, paleolongitude
dp, dm, axes of error ellipse in the polar and meridinal directions, respectively
AL, Assean Lake area; SL, Split Lake area; Gr, granites; all, all rock types

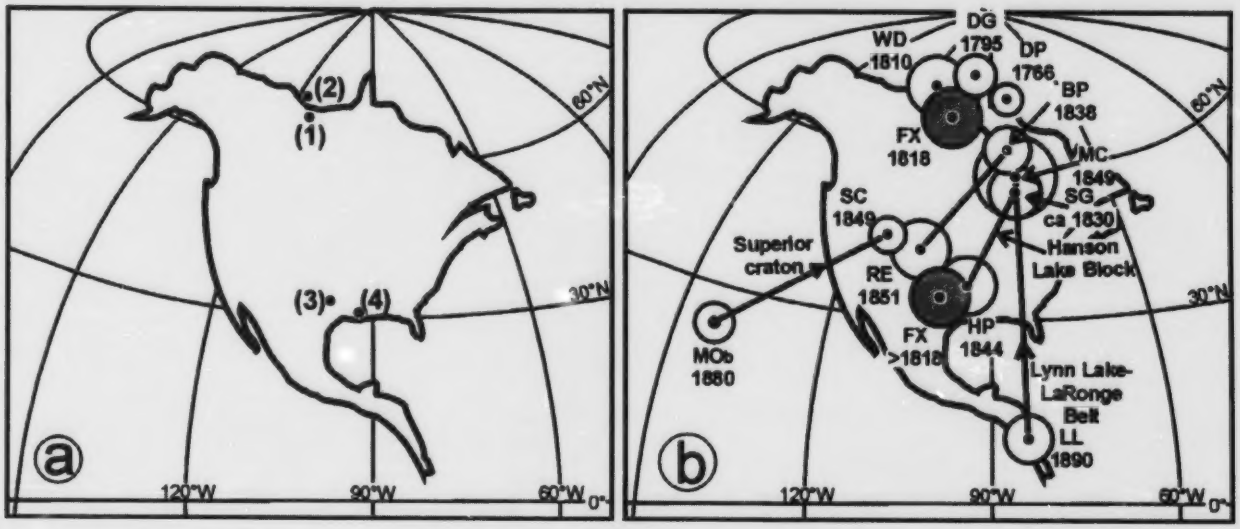


Figure GS-7-10: a) Locations of the paleopoles for the Fox Lake study (set numbers refer to Table GS-7-5). b) Apparent polar wander path (APWP) segments for the domains of the Superior craton and Trans-Hudson Orogen and the paleopole for the Fox Lake study. Abbreviations as in Figures GS-7-1 and -2.

the western part of the Fox Lake pluton indicates that conditions prevalent during metamorphism and deformation were at least garnet grade.

CONCLUSIONS

1) The paleomagnetic results from the more deformed part of the pluton along Awupak Narrows suggest that deformation along the Assean Lake Shear Zone occurred around 1810 Ma, in agreement with the average U/Pb titanite model age of 1818 Ma for the pluton. The tectonic conclusion would be that 1810 Ma represents rapid uplift and cooling and metamorphism from collision of the Superior craton with the adjacent Kisseynew Terrane on closure of the Trans-Hudson Orogen (THO). This paleomagnetic pole provides further support for the presence of a stillstand and hairpin in the THO apparent polar wander path between ca. 1830 Ma and 1770 Ma.
2) The location of the paleopole for the least-deformed part of the Fox Lake pluton along Split Lake suggests that the pluton's emplacement age is ca. 1850 Ma.
3) Al-in-hornblende geobarometry may be a viable tool for more than just granitoid rocks, as suggested by the internal consistency of the determinations in this study, values comparable to other determinations elsewhere, and the meeting of three requirements of the geobarometer by two mafic rock units.

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GS-8 GEOLOGY OF THE WESTERN PART OF THE FOX RIVER BELT (PARTS OF NTS 53M AND 53N)

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Peck, D.C., Potter, L., Desharnais, G., Scoates, R.F.J., Corkery, M.T. and Böhm, Ch.O. 2000: Geology of the western part of the Fox River Belt (parts of NTS 53M and 53N); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 38-41.

SUMMARY

This paper summarizes the results obtained from reconnaissance geological studies conducted during the 2000 mapping programs in the western part of the Fox River Belt. Critical new observations include:

1) The Lower Sedimentary formation and the adjacent Archean orthogneiss of the Superior Province host one or more suites of mafic (gabbronorite) intrusions that were not previously recognized.
2) Granitic rocks locally intrude the Lower Sedimentary formation in the southwestern part of the Fox River Belt, and may relate to previously unrecognized Proterozoic felsic plutonism that appears to have occurred along the Archean–Proterozoic boundary.
3) Paragneiss units occurring immediately to the north of the Fox River Belt are turbiditic greywacke that has been folded and metamorphosed at the amphibolite facies. Sampling was carried out on these units in order to determine their Nd-Sm model ages.
4) Limited sedimentological studies of the Middle Sedimentary formation and upper parts of the Lower Sedimentary formation suggest a strong similarity to the distal marine sedimentary rocks in the prospective parts of the Ospwagan Group in the Thompson Nickel Belt. In the Fox River Belt, most of the sedimentary rocks are undeformed and consist of repeated cycles, tens of metres thick, of early clastic (mudstone, siltstone, greywacke) and later chemical (carbonate overlain by sulphidic mudstone and/or sulphide-facies iron-formation) sedimentation.

INTRODUCTION

The metallogenic and tectonic evolution of the Fox River Belt (FRB) and adjoining gneissic terranes (Superior Province to the south and presumed Proterozoic metasedimentary gneiss belt to the north) is currently being revisited following more than a decade of inattention from government surveys and academe. Exploration activity during this same period was also minimal, and involved surface exploration, lake-bottom sediment geochemical sampling and a 10-hole diamond-drilling program carried out by Westminer Ltd. (now Western Mining International Ltd.) in 1992 and 1993. The stimulus for the recent, increased levels of geoscience programming in the FRB was provided by the minerals industry, in particular Falconbridge Limited and B. Dunlop, who began regional exploration of the FRB for Ni, Cu and platinum-group elements (PGE) in 1998. In 1999, the Manitoba Geological Survey and the University of Manitoba initiated field and geochemical investigations in the FRB (Peck et al., 1999) with the support of Falconbridge Limited. This year, two B.Sc. theses were completed by University of Manitoba students (M. Huminicki and C. Osiowy), and a Ph.D. study, focusing on the metallogeny and petrogenesis of ultramafic-mafic intrusions in the FRB, commenced this fall (G. Desharnais, University of Manitoba).

Most of our current knowledge of the geology of the FRB comes from a regional diamond-drilling program completed intermittently between the late 1950s and early 1970s by Inco Ltd. Previous field investigations of the FRB were conducted by the Manitoba Mines Branch, mainly between 1969 and 1977. The results of this work, together with a thorough petrological investigation of more than 100 Inco Ltd. drillholes, underpinned the only available regional maps (1:50 000 scale) and geological descriptions for the belt (Scoates, 1981, 1990). Most previous field studies were carried out in the Fox River valley, between the Gowan and Bigstone rivers (Fig. GS-8-1), an area characterized locally by semicontinuous bedrock exposure in and adjacent to the main river channel(s). Detailed mapping by previous workers was focused in the Great Falls area of the Fox River and along the Stupart River (Fig. GS-8-1). Geological and exploration data acquired for the FRB up to the mid-1980s are synthesized by Scoates (1990). Subsequent exploration of the FRB for PGE by BP-Selco (mid- to late 1980s) and a related M.Sc. study (Schwann, 1989) provided additional insights into the mineral potential and petrology of the largest known intrusion in the belt, the Fox River sill. Despite thorough studies of the core from approximately 140 diamond-drill holes in the FRB and local, detailed mapping along the Fox and Stupart rivers, the regional geology of the FRB remains poorly understood.

To redress this, the current regional geological program was designed with the following principal objectives:

1) Carry out reconnaissance, helicopter-borne, aerial surveys and follow-up surface investigations of previously unrecognized or undocumented outcrop areas.
2) Carry out detailed stratigraphic studies of volcanic and sedimentary formations in selected, well-exposed river outcrop areas (e.g. Syme et al., 1999);
3) Map and sample (for Nd-Sm isotopic analysis and possibly U-Pb detrital zircon geochronology) outcrops of paragneiss belonging to the Northern Gneiss Belt (age unknown) and orthogneiss belonging to the Southern Gneiss Belt (northern margin of the Superior Province; terminology *after* Scoates, 1981, 1990), in order to determine their age and tectonic significance.

The study area (Fig. GS-8-1) includes the western part of the FRB, adjacent parts of the Archean Superior Province to the south, and the southern part of the Northern Gneiss Belt. The centre of the study area is situated approximately 50 km south-southeast of the town of Gillam. The eastern boundary of the area is marked by the Stupart River, which also marks the eastern limit of outcrop in the FRB. Throughout the study area, bedrock outcrop is restricted to incised valleys that have been formed by the major rivers, which typically flow northeasterly toward Hudson Bay.

SYNOPTIC REGIONAL GEOLOGY

The Fox River Belt, as described by Scoates (1981, 1990), is an approximately 300 km long and 10 to 30 km wide, supracrustal sequence of approximately Proterozoic age that is intruded by coeval ultramafic and mafic sills and dykes. It is a significant part of the Superior Boundary Zone (SBZ), which separates Archean orthogneiss terranes and granite-greenstone belts of the Superior Province from arc and marine sedimentary-basin assemblages of the Trans-Hudson Orogen. The FRB represents a homoclinal, north-dipping and north-facing

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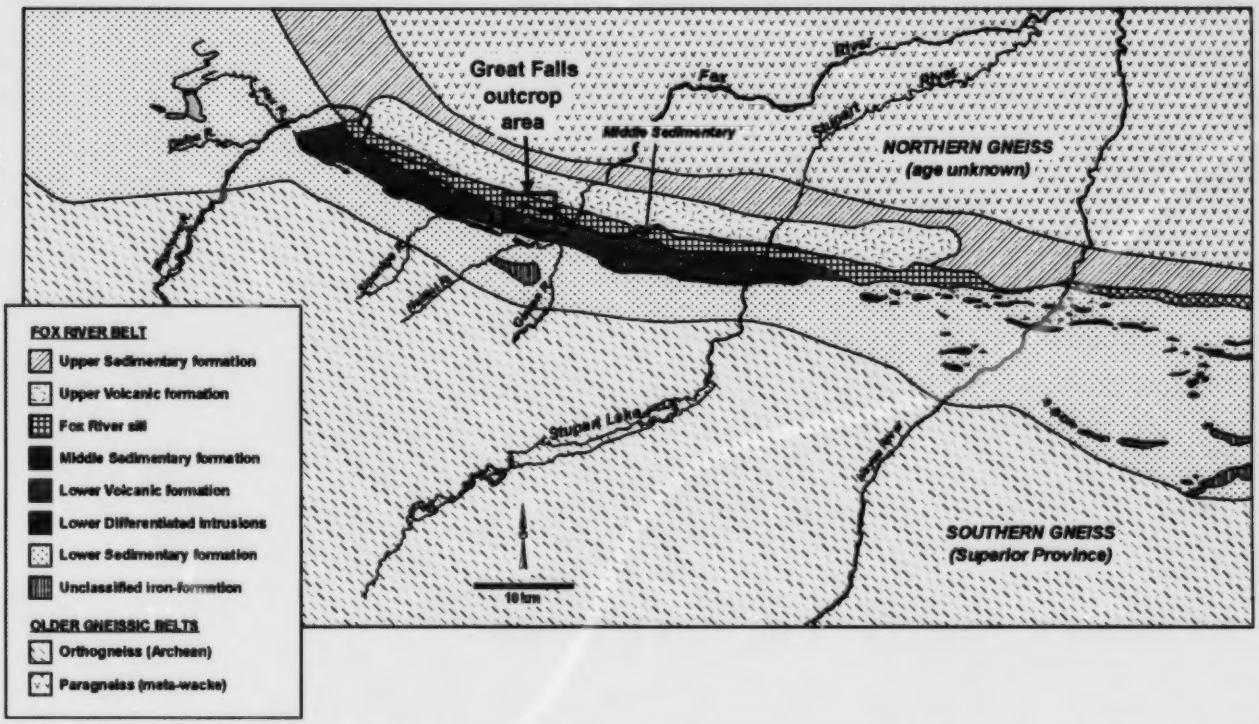

Figure GS-8-1: General geology of the western part of the Fox River Belt, showing the locations of reconnaissance mapping areas investigated in 2000 (modified from Scoates, 1981, 1990).

sequence that developed in a marginal rift during the Paleoproterozoic (ca. 1.9 Ga). The belt is relatively undeformed and exhibits low metamorphic grades (sub-greenschist to lower greenschist facies). It is likely the best preserved Paleoproterozoic rift basin sequence on Earth, contrasting with the complexly deformed rocks present in the Thompson Nickel Belt segment of the SBZ. Past and current exploration interest in the FRB is partially a reflection of the fact that it occurs in a regional metallotect (SBZ) for Ni-Cu-PGE deposits that hosts the Raglan camp in northern Quebec and the Thompson Nickel Belt in central Manitoba. The belt consists of three sedimentary rock formations, the Lower, Middle and Upper Sedimentary formations (Fig. GS-8-1; Scoates, 1981), that enclose an igneous domain comprising the Lower and Upper Volcanic formations and laterally continuous ultramafic-mafic intrusions that range from less than 50 m to more than 2.5 km in thickness. The largest intrusion in the FRB is the Fox River sill (*see* Scoates, 1990 for detailed descriptions). Smaller ultramafic to mafic intrusions (Lower Differentiated intrusions) occur in the northern part of the Lower Sedimentary formation.

REGIONAL FIELD PROGRAM – PRELIMINARY RESULTS

Although the general stratigraphy for the western part of the FRB (Scoates, 1981, 1990) is now well established, the lack of bedrock outcrop in the belt (<1% on average) has precluded a more detailed understanding of its geology and tectonic evolution. Nonetheless, many of the limited outcrop areas available for study are either not mapped or not studied in detail. Fieldwork was conducted during July and August 2000, concurrent with detailed mapping and prospecting in the Great Falls area on the Fox River (Desharnais et al., GS-9, this volume). The general locations of the outcrop areas studied during the current program are shown on Figure GS-8-1. Additional regional geological studies are planned for 2001, after which a revised 1:50 000 scale compilation map for the FRB will be prepared.

Gowan–Nettini Rivers Section

Two days were spent examining outcrops on the northern part of the Gowan River and the northeastern part of the Nettini River (Fig. GS-8-1). Bush outcrop located adjacent to the Nettini River is situated within the Lower Volcanic formation. Plagioclase-phyric massive flows of the upper part of the Lower Volcanic formation are present on the Gowan River, north of the mouth of the Nettini River. The flows are up to 25 m thick, have a well developed lower colonnade and upper entablature (*see* Syme et al. (1999) for relevant, detailed volcanological descriptions for the FRB). One of the flows has a distinctive, upper, rubbly flow-top that has been affected by shearing along the contact between the top of this flow and the base of the overlying flow. The plagioclase phenocrysts are interpreted to represent xenocrysts entrained from a subvolcanic, differentiated magma chamber. They display strong resorption by the enclosing basaltic matrix and are up to 1.5 cm in length. These flows are plagioclase-phyric throughout and have an overall basaltic composition.

Outcrops in and adjacent to the Gowan River consist of massive and pillowed flows of basalt and komatiitic basalt. Massive flows appear to predominate in the north, proximal to the confluence of the Gowan and Fox rivers, whereas pillowed flows are more prevalent at and south of the northeastern termination of the Nettini River (Fig. GS-8-1). The volcanic architecture and flow regimes appear to be similar to those described for the Lower Volcanic formation outcrop areas along the Fox River (Scoates, 1981; Syme et al., 1999).

A single, large outcrop of the Middle Sedimentary formation occurs less than 1 km south of the confluence of the Gowan and Fox rivers. The outcrop comprises thinly bedded, turbiditic mudstone-siltstone-wacke sequences but, unlike most of the known Middle Sedimentary formation outcrops, does not appear to contain pyritic mudstone (Scoates, 1990). No outcrops of the Fox River sill or any Lower intrusions were recognized.

Northern Gneiss Belt

Part of one day was spent examining river and bush outcrop of Northern Gneiss Belt rocks in the Fox River, approximately 500 m north of the inferred northern contact of the Fox River Belt (Fig. GS-8-1). Several samples were collected from the paragneiss units that were observed in these outcrops. These samples will be used to study the Nd-Sm model ages for the rocks, an extension of the work being done by Ch.O. Böhm, M.T. Corkery and L.M. Heaman (University of Alberta) on the geochronology and tectonic evolution of the northwestern part of the Superior Boundary Zone (Corkery et al., GS-25, this volume). The Northern Gneiss Belt outcrops investigated are interpreted as representing distal marine turbiditic greywacke and display planar bedding in the form of decimetre-scale, alternating gritty and muddy layers. The rocks are folded about south-dipping axial planes and appear to have been recrystallized at mid-amphibolite facies, which promoted minor incipient melting of pelitic components. However, their age and tectonic relevance remains equivocal. They could represent Archean paragneiss, perhaps similar to those rocks occurring north of the Assean Lake Fault and between the FRB and the Thompson Nickel Belt segments of the SBZ (e.g., Corkery et al., GS-25, this volume). Alternatively, these gneiss units could represent Paleoproterozoic ocean-floor sequences analogous to parts of the Trans-Hudson Orogen in northern and western Manitoba (Kisseynew Domain). If the Nd-Sm work is successful, detrital zircon U-Pb geochronology may be undertaken to obtain more precise depositional ages for the rocks.

Southern Gneiss Belt

The Southern Gneiss Belt rocks have now been observed and sampled at three locations within the study area: 1) south Stupart River, 2) southern fork of the Sipanigo River, and 3) Bigstone River. These rocks do not appear to have ever attained granulite facies grade, and are therefore unlikely to be part of the Pikwitonei Domain, which flanks or is part of the northwestern segment of the Superior Boundary Zone. The metamorphic grade is estimated, from petrology, to be middle amphibolite facies and is therefore similar to the grade observed in the Northern Gneiss Belt. At the Bigstone River, a massive granodiorite pluton can be followed in outcrop along the river for at least 8 km to the south of the inferred contact with the Fox River Belt. The contact appears to be exposed on the Bigstone River, and seems to involve a planar, unfaulted contact between the granodiorite to the south (?Archean) and a gabbro±sedimentary sequence to the north (base of the Fox River Belt). This outcrop area requires further study when water levels are lower.

Fox River South

Scattered outcrops occur in the bush and in the river within the southern part of the Fox River valley, immediately west of Great Falls and continuing west for up to 20 km. These outcrops expose the Lower Volcanic formation and, rarely, the Middle Sedimentary formation (Scoates, 1981, 1990). A forest fire (August 2000) affected many of these outcrops, which should therefore be more amenable to detailed mapping following the spring runoff in 2001. Accordingly, detailed investigations of these outcrops has been deferred until next year.

Gabbroic Intrusions in the Southern Part of the Fox River Belt

Based on outcrops inspected during the 1999 and 2000 field programs, it is believed that there are two or more suites of east-striking (080–100°) mafic intrusions that occur on either side of the contact between the Southern Gneiss Belt and the FRB. One of these occurs within the southern part of the FRB (Bigstone gabbro) and is likely hosted by the lower part of the Lower Sedimentary formation. The other occurs in the Southern Gneiss Belt, which represents the northernmost part of the Superior Province, immediately to the south of the FRB (Stupart dykes).

The Bigstone gabbro refers to gabbronorite bodies of presently unknown dimensions that are exposed in outcrops on the Bigstone River (Fig. GS-8-1). These exposures occur approximately 0.5 to 1 km north of the indicated contact (Scoates, 1990) between the Southern Gneiss Belt and the Fox River Belt. The Bigstone gabbro may also include gabbroic bodies recognized by staff of Falconbridge Ltd. in 1999, during reconnaissance studies along the Sipanigo River, several kilometres southwest of its junction with the Fox River (Fig. GS-8-1). All of these gabbro bodies have been metamorphosed at greenschist facies grade and appear to be hosted by the lower to middle parts of the Lower Sedimentary formation. They are distinguished from more magnesian intrusions (Lower Differentiated intrusions) that have been intersected in diamond-drill holes in the upper part of the Lower Sedimentary formation (Scoates, 1990). Additional, more detailed investigations of the gabbroic rocks exposed in the Bigstone River are planned for the 2001 field season.

Last year, a suite of easterly trending, layered gabbroic dykes was recognized in bush outcrops occurring within Archean basement gneiss, approximately 2 km south of the southern boundary of the Fox River Belt and adjacent to the Stupart River (Fig. GS-8-1). These intrusions, emplaced into partially migmatized tonalitic gneiss and derived granodiorite intrusions, are referred to as the Stupart dykes. This year, another mafic dyke was recognized in approximately the same position relative to the southern margin of the FRB, along the south fork of the Sipanigo River (Fig. GS-8-1). This dyke was also emplaced into tonalitic gneiss and is tentatively correlated with the Stupart dykes. In 1999, a sample of a pegmatitic gabbro/diorite pod was obtained (DCP) for U-Pb zircon geochronology (L. Heaman, University of Alberta) from an approximately 40 m thick, well layered mafic dyke occurring in the Southern Gneiss Belt (Superior Province), adjacent to the Stupart River. Age determinations were not available at the time this report was written.

These recent observations may be significant from both a tectonic and metallogenic viewpoint. Previously, it was believed that the first major intrusions emplaced into the FRB were the Lower Differentiated intrusions. The existence of the Bigstone gabbro and Stupart dykes indicates that the volume of mafic and ultramafic intrusions in the FRB is much larger than previously believed, and that the entire Lower Sedimentary formation may be permeated with sills and dykes of mafic to ultramafic composition. All of these intrusions should be explored for Ni, Cu and platinum-group element mineralization. The Stupart dykes may be coeval with mafic and ultramafic intrusions in the FRB and are parallel to stratigraphic layering in the FRB. In this way, the Stupart dykes are analogous to the Molson dykes (Scoates and Macek, 1978), most of which are parallel to the Thompson Nickel Belt and coeval with ultramafic bodies in that belt.

GENERAL OBSERVATIONS

The western part of the FRB is the only part of the belt where it is possible to collect critical field data. Observations made during the current and previous projects provide key constraints on both the metallogenic and tectonic evolution of the FRB. For this reason and because of the paucity of outcrop throughout the belt, it is important to continue with detailed studies of available outcrop areas. This work should include detailed mapping, lithogeochemical sampling, structural studies and, where appropriate, geochronological studies. Field observations will help to constrain and refine lithogeophysical mapping (based mainly on airborne magnetic data) of the unexposed parts of the FRB, which represents the only practical method of regional correlation that can be applied to the FRB. This approach must also rely heavily on drilling results. To date, more than 140 diamond-drill holes have been completed in the FRB. Some of the more recent drillholes have been relogged as part of the current program. Some key, initial results include the following:

1) The upper part of the Lower Sedimentary formation comprises repeated cycles of marine sedimentation. Each cycle consists of a

lower, clastic mudstone-siltstone sequence and an upper, thinner, chemical sedimentary sequence that includes a lowermost carbonate and/or calc-silicate unit and an overlying sequence of iron-formation units and sulphidic mudstone (locally graphitic). These sedimentary cycles are typical of distal marine basins and may mark the change from passive (clastic sedimentation) to active (chemical sedimentation) rifting in the FRB. They are also strikingly similar in their internal stratigraphy to the Ospwagan Group in the Thompson Nickel Belt, which hosts most of the most valuable Ni deposits in that belt.

2) A series of ultramafic and mafic intrusions occurs between the main Lower intrusions and the base of the Lower Sedimentary formation. These intrusions were intersected by Westminer Ltd. drillholes (1993). Their position in the FRB suggests that the Lower Volcanic formation is, at least locally, thinner than was previously believed, and that there are several levels of emplacement for these intrusions.

The large body of massive to weakly foliated granite to granodiorite exposed on the Bigstone River may signify a syn- or post-rifting period of felsic plutonism in the FRB and the adjacent part of the Superior Province. A sample of this granite was collected for geochemical analysis and possible U-Pb geochronology. Similar granitic rocks were observed in a boudinaged dyke emplaced into sheared metasedimentary rocks along the Dafoe River, several kilometres west of the Bigstone River and presumably in the southern part of the Lower Sedimentary formation.

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GS-9

GEOLOGY AND MINERAL OCCURRENCES OF THE FOX RIVER SILL IN THE GREAT FALLS AREA, FOX RIVER BELT (PART OF NTS 53M/16)

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Desharnais, G., Peck, D.C., Theyer, P., Potter, L., Huminicki, M., Scoates, R.F.J., Halden, N.M. and Kohut, G., 2000: Geology and mineral occurrences of the Fox River sill in the Great Falls area, Fox River Belt (part of NTS 53M/16); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 42-48.

SUMMARY

The Great Falls area provides the best known exposures of the Fox River sill within the Fox River Belt. Mapping completed in 2000 focused on outcrops of the Marginal zone and enabled completion of a 1:1000 scale geological map of the northern half of the Great Falls outcrop area. Mapping has shown that the lower part of the Fox River sill in this area is largely undeformed, with the exception of an orthogonal set of subvertical, brittle faults that have developed along and perpendicular to major petrological boundaries. These faults are believed to have developed contemporaneously with active rifting in the Fox River Belt (peak extension). In the Great Falls study area, the lower part of the Fox River sill has a chilled contact with older pyritic mudstone and siltstone of the Middle Sedimentary formation. These sedimentary rocks are typically hornfelsed within the several metre thick contact aureole that is developed adjacent to the southern margin of the sill. The Marginal zone of the Fox River sill comprises a lower, reverse differentiated, mafic to ultramafic unit (Basal Contact unit) and two normally differentiated ultramafic to mafic cyclic units.

Surface prospecting in 1999 and 2000 delineated disseminated Fe-Cu-Ni sulphide mineralization within the Marginal zone at three different stratigraphic levels: 1) in the Basal Contact unit, immediately north of the Middle Sedimentary formation; 2) in pyroxene-rich ultramafic rocks developed along the base of the second cyclic unit (KO zone); and 3) in the gabbroic upper part of the second cyclic unit. The KO zone is a stratabound zone of sulphide mineralization, enriched in Cu, Ni and platinum-group elements (PGE), that developed immediately above the irregular contact between the host ultramafic rocks of the second cyclic unit (UM2 subunit) and the upper, leucogabbroic part of the first cyclic unit (LG1 subunit). The irregular form of the LG1-UM2 contact is interpreted to have developed by erosion of the older UM1 subunit through current action associated with the emplacement of the second cyclic unit (UM2-LG2). Recent analytical results confirm the PGE-rich nature of the KO zone, which has a maximum grade of 5.4 g/t combined Pd+Pt+Au, 2.3% Cu and 1.1% Ni, based on a limited number of surface grab samples. Total sulphide abundance in the KO zone increases toward the base of the UM2 subunit, where the sulphide minerals occur within a pyroxenite layer that ranges from less than 1 to 3 m in thickness. The mineralization appears to reach its maximum abundance in the numerous, small, centimetre- to metre-wide troughs that are developed along the LG1-UM2 contact. However, disseminated sulphide mineralization of presently unknown grade also extends northward into overlying lherzolite and olivine pyroxenite, so that the total thickness of the KO zone is not presently known. Irregularly disseminated sulphide mineralization is also present in the gabbroic upper part of the second cyclic unit (LG2 subunit) but appears to have low base-metal and PGE contents. Poddy, metre-size leucogabbro-leucodiorite bodies are locally present within the UM2 subunit at a distance of 1 to 5 m above its base. These pods are also locally mineralized and are interpreted to represent melts derived from the LG1 subunit that entrained pre-existing sulphide minerals in the KO zone.

The KO zone is continuous along strike within the Great Falls area (approx. 1.5 km of strike length) and, given the lateral continuity of the Fox River sill stratigraphy on a regional scale, is likely to be present in other parts of the Fox River Belt. The host pyroxenite layer exhibits normal size and compositional grading of both sulphide and silicate components, indicating derivation through settling of immiscible sulphide liquid and pyroxene from a S-saturated ultramafic magma. Therefore, it is suggested that the KO zone represents a basal-segregation type of magmatic sulphide mineralization. If this is correct, then the high Cu:Ni ratios and Pd+Pt abundances in the KO zone are difficult to reconcile with the primitive composition of the host rocks. Accordingly, it is likely that more Ni-rich sulphide mineralization, which would be expected to develop in the Mg- and Ni-rich ultramafic parent magmas to the UM2 subunit, may be present in larger troughs/embayments along the LG1-UM2 contact. These preliminary findings, coupled with previous investigations of stratabound sulphide mineralization in the Upper Central Layered zone, demonstrate the significant potential for magmatic Ni-Cu-PGE sulphide deposits in the Fox River sill.

INTRODUCTION

The regional geology of the Fox River Belt (FRB) is discussed by Scoates (1981, 1990). Results obtained from previous mapping programs in the Great Falls area (Fig. GS-9-1) are given in Scoates (1981, 1990), Peck et al. (1999) and Syme et al. (1999). The FRB is an east-trending supracrustal belt, approximately 300 km long and 30 km wide, comprising abundant mafic and ultramafic intrusions (including the approx. 2 km thick Fox River sill), submarine flow sequences, and marine clastic and chemical sedimentary rocks. The FRB represents the best preserved (least deformed) part of the Superior Boundary Zone, which extends to the northeast through Hudson Bay into the Cape Smith Belt in northern Quebec, and to the southwest into the Thompson Nickel Belt in central Manitoba. The FRB is poorly exposed, but limited geological mapping, combined with detailed drill core studies described by Scoates (1981, 1990), has shown that it is a well preserved, Proterozoic, rifted continental-margin sequence. Heaman et al. (1986) have dated the Fox River sill (Marginal zone) at 1883 Ma using the U-Pb method. Results from the current program, which focused on the geology and mineral occurrences of the Marginal zone in the Great Falls area, are discussed below. Preliminary lithological sections for the Great Falls area are given in Figure GS-9-2. This year's mapping results are shown in detail in Peck et al. (2000).

The objectives of the 2000 field program in the Great Falls area were to:

1) complete a detailed 1:1000 scale geological map of the northern part of the Great Falls outcrop area; and
2) document the mineral occurrences throughout the study area, with an emphasis on the Marginal zone, in which a PGE-Cu-Ni occurrence (MZ1 showing) was identified in June of 1999 (Peck et al., 1999).

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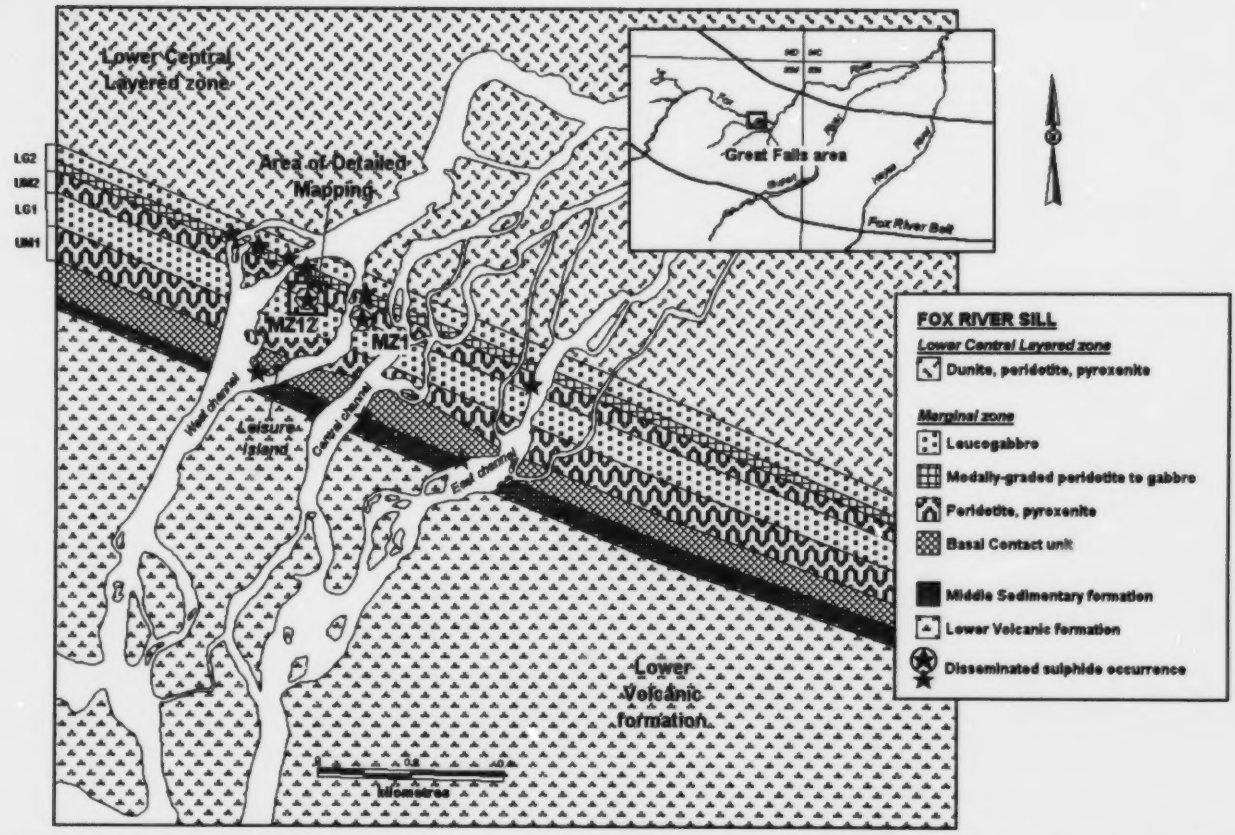

Figure GS-9-1: General geology of the Great Falls outcrop area, Fox River Belt. Inset map shows location of the study area in the western part of the Fox River Belt. Box outlines the Leisure island detailed mapping area.

GEOLOGY OF THE FOX RIVER SILL IN THE GREAT FALLS AREA

The Great Falls provides the best outcrop exposure of the Fox River sill in the FRB. It covers an area of about 2.1 by 2.8 km (Fig. GS-9-1) that is underlain by a homoclinal, north-facing, conformable, supracrustal and intrusive sequence comprising (from south to north) the upper part of the Lower Volcanic formation, the Middle Sedimentary formation, and the Marginal zone and lower part of the Lower Central Layered zone of the Fox River sill (Fig. GS-9-1). Peak metamorphism in the area is low grade, and varies between the prehnite-pumpellyite and lower greenschist facies (Scoates, 1990). Primary mineralogy is typically well preserved, particularly in the rocks of the Marginal zone (Huminicki, 2000). In this study, no attempt was made to distinguish between ortho- and clinopyroxene in the field, so the generic term 'pyroxenite' (which could represent clinopyroxenite, websterite or orthopyroxenite) is used. Similarly, the field term 'gabbro' may represent norite, gabbronorite or gabbro. The more precise terminology is advanced in the discussion of the geochemical results.

The study area is relatively undeformed, with the exception of an orthogonal pair of prominent fractures and/or brittle faults. One of these is coplanar with igneous layering (e.g. striking between 110 and 115° and having vertical or near-vertical dips), and the other cuts layering at approximately right angles, trending at approximately 020° (*see* Peck et al., 2000). The latter structures locally show north-northeast-directed lateral displacement of the stratigraphy. Within the study area, this north-northeast apparent horizontal displacement is largely taken up on a single major fault that has a dextral offset of layering of approximately 200 m (Peck et al., 2000). Some of the units in the Fox River sill appear to thicken across this major fault, suggesting that it could represent a syn-magmatic structure, coincident with peak extension in the Fox River Belt during an active rifting stage. Locally, a south-side-up sense of vertical displacement is evident on the east-southeast-trending (layer-parallel) faults, suggesting that they are normal faults. The Lower Volcanic formation in the Great Falls area is composed of a layered sequence of pillowed and massive, compound flows of basaltic to komatiitic-basaltic composition (*see* Syme et al., 1999). In the study area, the Middle Sedimentary formation forms a thin (generally <100 m), poorly exposed, discontinuous package of laminated, pyritic mudstone and siltstone (Fig. GS-9-1). These sedimentary rocks are hornfelsed near the contact with the sill, where their primary bedding has been replaced by a massive, aphanitic texture.

At Great Falls, the Marginal zone comprises three main subunits, each approximately 100 m thick, including (from south to north) the Basal Contact unit, Cyclic unit 1 and Cyclic unit 2 (Fig. GS-9-1, -2). Most of the ultramafic and mafic rocks in the study area are cumulate rocks made up of variable proportions of cumulus olivine, clinopyroxene, chromite (minor) and postcumulus (poikilitic) orthopyroxene (ultramafic units), or cumulus clinopyroxene, orthopyroxene and plagioclase (mafic units). The sense of differentiation of the units is such that the stratigraphic younging direction is assumed to be to the north, in keeping with flow tops measured in the underlying Lower Volcanic formation. The Basal Contact unit locally contains hornfelsed xenoliths of the Middle Sedimentary formation and minor disseminated pyrrhotite±chalcopyrite at its base. The Basal Contact unit is overlain by Cyclic unit 1, which is normally differentiated and 100 to 150 m thick. Cyclic unit 1 grades abruptly from a lower lherzolite and olivine pyroxenite subunit (UM1 subunit) to a more leucocratic, melagabbro (minor) to gabbro to leucogabbro sequence (LG1 subunit). An anorthosite layer, less than 0.5 m thick, is locally present at the top of the LG1 subunit, and

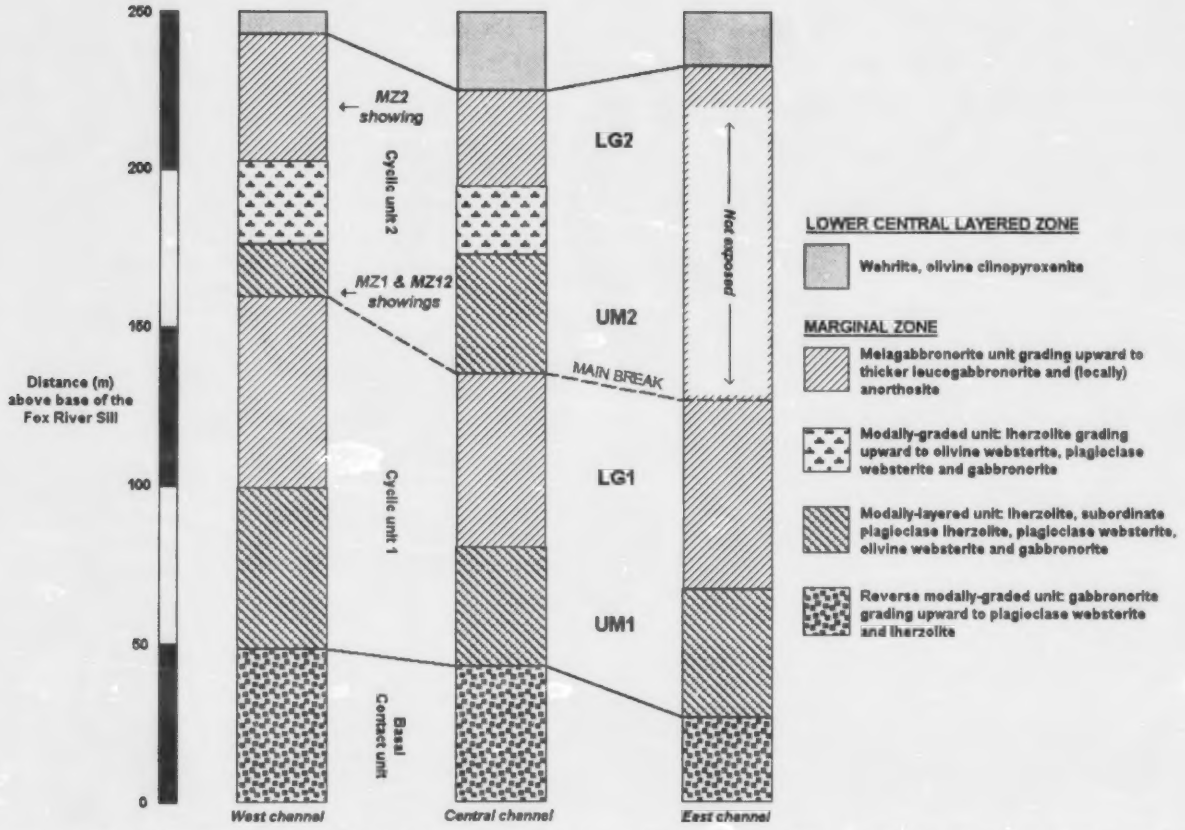

Figure GS-9-2: Preliminary lithological sections for the West, Central and East channels of the Fox River in the Great Falls area (modified from Scoates, 1990; Peck et al., 1999).

leucotonalite veins are commonly present in the upper 20 m of this subunit. Cyclic unit 1 and Cyclic unit 2 are separated by the Main break (Fig. GS-9-1, -2), a decimetre- to metre-wide break in outcrop that forms a prominent east-southeast-trending lineament on aerial photographs. The Main break locally appears to represent a vertical (normal) fault, but elsewhere is simply a narrow fracture represented by a 1 to 2 m wide break in the outcrop that reflects preferential weathering of the less resistant ultramafic rocks of the lower parts of the UM2 subunit relative to the more resistant gabbroic rocks of the LG1 subunit. Cyclic unit 2 is 100 to 150 m thick and interpreted to have formed from a single pulse of magma emplaced before the rocks in Cyclic unit 1 were completely solidified. Field observations, not yet corroborated by petrographic studies, indicate a crystallization order for Cyclic unit 2 of orthopyroxene→olivine→chromite→clinopyroxene→plagioclase, which is atypical for the Marginal zone and for the Fox River sill in general (olivine→chromite→clinopyroxene→plagioclase→orthopyroxene; Scoates, 1990). Cyclic unit 2 is further distinguished by the persistent development of disseminated pyrrhotite±chalcopyrite mineralization, which suggests that it, unlike Cyclic unit 1, crystallized from sulphur-saturated magma.

The lowermost 1 to 3 m of the UM2 subunit host several occurrences of chalcopyrite-dominant, disseminated sulphide mineralization (Fig. GS-9-1). In the Leisure island (unofficial name) detailed-mapping area (Fig. GS-9-1), these sulphide minerals are principally confined to a coarse-grained pyroxenite layer and appear to be best concentrated within decimetre- to metre-size troughs that are erratically developed along the LG1-UM2 contact (Fig. GS-9-3, -4). The sulphide mineralization at the base of the UM2 subunit is referred to as the KO zone. Chalcopyrite is the dominant sulphide mineral, and attains a maximum of 15% near the base of the troughs. The unit also contains lesser amounts of pyrrhotite (up to 8%) and pentlandite, whose presence has been inferred from chemistry. Generally, the sulphide minerals are segregated into a lowermost, coarse-grained, pyrrhotite-rich horizon (approx. 10 cm), which grades upward into a wider, finer grained, chalcopyrite-dominant layer (up to 3 m thick but typically <1 m). The total sulphide content generally decreases upward.

At the original discovery outcrop (MZ1 showing, Fig. GS-9-1; *see also* Peck et al., 1999), the pyroxenite layer is absent but the mineralization persists and is contained within a pod-shaped body of leucogabbro that becomes more pyroxene rich and coarser grained toward its base. Where present, the pyroxenite layer varies from very coarse grained (crystals up to 3 cm long) to medium grained (crystals <0.5 cm long). The local absence of the pyroxenite layer is considered to represent preferential erosion of talc-amphibole–altered parts of the layer, rather than lateral thickness variations in the layer. The basal UM2 pyroxenite grades abruptly upward into lherzolite and olivine pyroxenite. The lherzolite is a fine-grained olivine cumulate that also contains coarse-grained orthopyroxene oikocrysts and finer grained, subequant clinopyroxene. At Leisure island (Fig. GS-9-1), no mineralization was observed in the olivine cumulates overlying the mineralized pyroxenite layer, but the olivine cumulates are not well exposed. Elsewhere, such as at the MZ1 showing (Fig. GS-9-1), centimetre-thick bands of disseminated sulphide-bearing lherzolite are recognized within 1 to 3 m of the Main break, in areas where the basal pyroxenite layer is not observed. At the MZ1 showing, chalcopyrite-rich disseminated sulphide mineralization occurs in a pod-shaped, variably textured, and differentiated body of gabbro to leucogabbro that is interpreted to represent melts derived from the underlying LG1 subunit. Similar gabbroic pods, lacking abundant

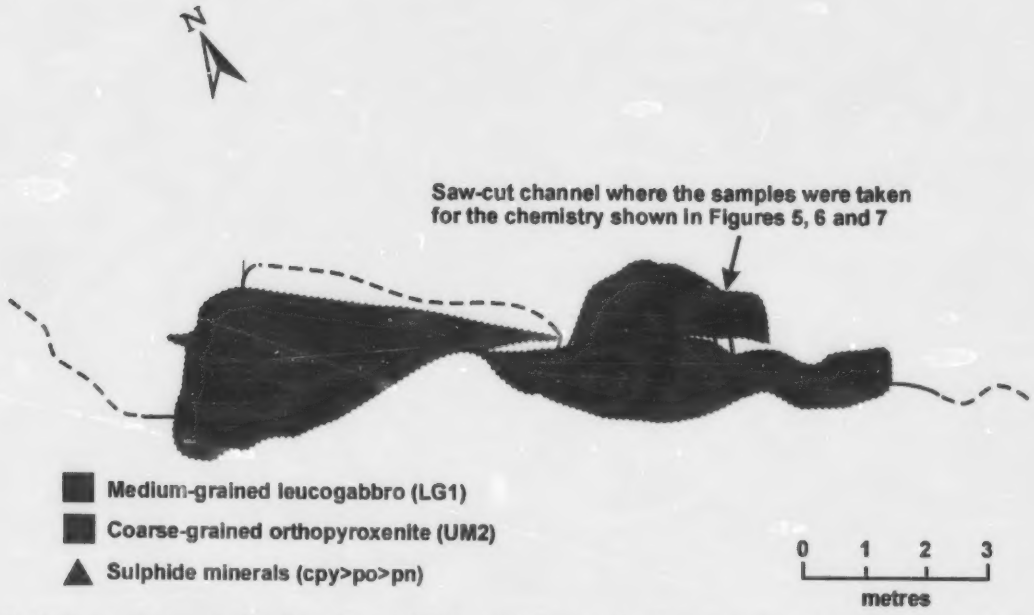
N
Saw-cut channel where the samples were taken
for the chemistry shown in Figures 5, 6 and 7
Medium-grained leucogabbro (LG1)
Coarse-grained orthopyroxenite (UM2)
Sulphide minerals (cpy>po>pn)
0 1 2 3
metres

Figure GS-9-3: Geology and sample locations for the MZ12 sulphide showing outcrop (Leisure island, Great Falls; Fig. GS-9-1), illustrating the irregular, scalloped contact between the LG1 and UM2 subunits of the Marginal zone, Fox River sill.

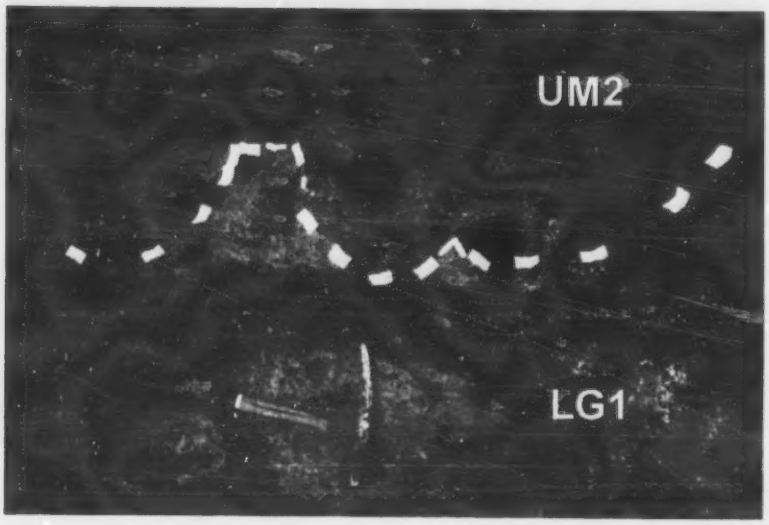

Figure GS-9-4: Photograph of the scalloped contact between LG1 and UM2 units; the contact has been outlined with pieces of white tape.

sulphide minerals, are present along most of the exposed parts of the lower UM2 subunit. The UM2 subunit grades upward into a melagabbro layer, several metres thick, that in turn grades upward into the LG2 subunit (gabbro and leucogabbro). Trace to 15% disseminated, pyrrhotite-rich sulphide mineralization is erratically distributed throughout both the transitional melagabbro layer and the LG2 subunit. Cyclic unit 2 is overlain by a pyroxenite layer representing the base of the approximately 1 km thick Lower Central Layered zone (Scoates, 1990).

THE KO ZONE: A NEW CU-NI-PGE TARGET IN THE MARGINAL ZONE OF THE FOX RIVER SILL

Mapping and surface prospecting in the Marginal zone of the Fox River sill, completed during the 1999 and 2000 field programs, delineated several new sulphide showings in the lower part of the UM2 subunit. The initial discovery (MZ1 showing) was made by K. Olshefsky of Falconbridge Ltd. in July of 1999. In recognition of this, the Cu-Ni-PGE-enriched parts of the UM2 subunit, comprising the basal pyroxenite layer and the sulphide-bearing parts of the overlying lherzolite–olivine pyroxenite sequence, are herein referred to as the KO zone.

Geological Characteristics and Genetic Constraints of KO Zone Mineralization

The KO zone mineralization, typically restricted to the lowermost 1 to 3 m of the UM2 subunit, comprises medium- to coarse-grained, disseminated pyrrhotite and chalcopyrite (the presence of pentlandite has not been visually confirmed but is inferred from assays; Fig. GS-9-3, -4). The sulphide minerals display a range of textures, including interstitial-irregular, net-textured and blebby. The following is a summary of the principal geological attributes of the KO zone, based on field observations made during the 1999 and 2000 field programs and research conducted by Huminicki (2000). In addition, preliminary genetic interpretations are provided where constrained by the field data. A graduate thesis study, recently initiated by the senior author at the University of

Manitoba, will further consider the metallogeny of the KO zone.

1) The KO zone mineralization always occurs on or immediately above the LG1-UM2 contact, and is clearly stratabound and locally stratiform in nature, as reflected by the presence of thin sulphide-rich layers at the MZ1 showing. These layers are conformable with igneous layering in the host lherzolite–olivine pyroxenite sequence.
2) The mineralization is concentrated within small-scale basin structures that are interpreted as representing primary topographic features caused by scouring of the LG1 subunit by the UM2 magma.
3) The development of small-scale basin structures in the Great Falls outcrop area, where the Marginal zone internal subunit contacts are extremely planar on the scale of the map area, indicates that larger, second-order (e.g. decimetre-scale) and first-order (kilometre-scale) structures could exist along strike to the west and east.
4) Where most concentrated, the KO zone sulphide mineralization displays textural and compositional features reflecting effective separation of immiscible, magmatic sulphide liquid from a S-saturated ultramafic parent magma, including normal size and compositional grading of both sulphide and silicate components. Accordingly, the mineralization may be better concentrated in larger basin structures that, to date, have not been seen in outcrop or drill core.
5) The KO zone is principally hosted by an isomodal, pyroxene-rich layer, but is also locally present in overlying lherzolite and olivine pyroxenite that make up most of the approximately 50 m thick UM2 subunit. Additional surface prospecting and, ultimately, drilling will be required to determine the dimensions of the zone.
6) Leucogabbroic pods, up to 10 m long and 5 m thick, occur within several metres of the LG1-UM2 contact in the UM2 subunit. The pods display variable textures (fine grained to pegmatitic) and compositions (melagabbro to leucodiorite) and, like the pod at the MZ1 showing (Peck et al., 1999; Huminicki, 2000), are locally mineralized and normally differentiated with respect to their grain size (fining upward) and mineralogy (increase in plagioclase content upward). Based on the existing mapping, the pods are believed to represent small volumes of buoyant, evolved gabbroic magma that originated in the underlying LG1 subunit. These gabbroic melts could have been generated by reheating and melting of the top of the LG1 subunit due to heat transfer from the overlying UM2 magma. No pods have been recognized in the middle or upper parts of the UM2 subunit, and it is not known why the pods failed to ascend farther above the UM2 contact. Their origins will be investigated in more detail by the senior author as part of the previously mentioned graduate thesis project.

Preliminary Geochemical Results

Assay and whole-rock analytical data have been obtained for 40 samples collected during the 2000 field season from the Leisure island exposures of the KO zone mineralization (Table GS-9-1). Analytical work was carried out by Activation Laboratories, Ancaster, Ontario; QA/QC data are available on request. Additional data, acquired during the 1999 field season from other outcrops of the UM2 and LG2 subunits in the Great Falls area, are discussed in Peck et al. (1999) and Huminicki (2000). Collectively, these data provide additional constraints on the genesis of the mineralization, including the following critical features:

1) Collectively, the UM1-LG1 subunits represent a normally differentiated sequence of lherzolite→websterite→melagabbronorite→gabbronorite→leucogabbronorite→anorthosite, in which Al content increases and Mg content decreases up section (Huminicki, 2000).Note that both geochemical and petrographic data are used here to support further subdivision of the various gabbroic and ultramafic rocks, based on the relative abundances of clino- and orthopyroxene.
2) The UM2 subunit is very different in chemical composition compared to most of the other ultramafic intrusive and volcanic units in the Fox River Belt. This reflects the atypically high orthopyroxene and sulphide mineral content in the UM2 subunit.
3) The base of the UM2 subunit shows the first significant increase in S and chalcophile metal content in the Fox River sill, with the notable exception of the disseminated sulphide mineralization recognized in

Table GS-9-1: Geochemistry of selected samples collected from MZ12 showing. Oxides expressed in per cent, trace elements in ppm.

Sample No.	M1B-A	M1B-B	M1B-C	M1B-D	M1B-E	M1B-F	M1B-G	M1B-H	M1B-I	M1B-J	M1B-K	M1B-L	M1B-M	M2EA
SiO_2	49.69	46.10	44.77	46.54	44.54	45.35	45.94	49.62	47.16	46.70	44.97	47.10	46.28	49.36
Al_2O_3	16.16	8.67	9.41	7.91	9.42	8.58	7.77	6.08	7.22	7.85	9.66	7.72	8.08	4.42
Fe_2O_3	7.99	14.90	15.22	14.28	15.00	14.44	14.96	11.36	12.97	12.80	14.12	12.90	13.07	15.47
MnO	0.14	0.18	0.18	0.18	0.18	0.17	0.17	0.18	0.18	0.15	0.19	0.18	0.18	0.15
MgO	11.49	20.57	21.81	21.40	22.18	21.70	21.25	21.17	21.59	22.57	22.14	22.17	22.03	20.22
CaO	12.83	9.15	8.38	9.18	8.26	9.10	9.11	10.85	10.27	9.43	8.45	9.50	9.89	9.70
Na_2O_3	1.08	0.22	0.11	0.13	0.13	0.22	0.24	0.28	0.24	0.17	0.17	0.19	0.18	0.16
K_2O	0.37	-0.01	0.06	0.07	0.02	0.07	0.07	0.08	0.09	0.04	0.03	0.02	0.04	0.07
TiO_2	0.25	0.19	0.24	0.29	0.25	0.33	0.45	0.36	0.26	0.25	0.28	0.21	0.24	0.43
Cr	145	211	846	866	1040	991	1030	564	732	899	143	662	866	315
Co	37	103	123	122	137	120	116	107	83	87	50	45	69	225
Ni	215	2040	1980	1830	1880	1690	1730	1520	895	663	311	461	635	9950
Cu	52	3400	3760	6980	3910	6590	11800	302	176	283	42	119	143	20700
Zn	64	65	62	57	84	62	72	41	82	52	53	48	66	78

M1B-A: gabbro at southern end of saw cut shown on Figure GS-9-3 (subunit LG1)
M1B-B: orthopyroxenite above M1B-A (UM2)
M1B-C: orthopyroxenite above M1B-B (subunit UM2)
M1B-D: orthopyroxenite above M1B-C (subunit UM2)
M1B-E: orthopyroxenite above M1B-D (subunit UM2)
M1B-F: orthopyroxenite above M1B-E (subunit UM2)
M1B-G: orthopyroxenite above M1B-F (subunit UM2)
M1B-H: orthopyroxenite above M1B-G (subunit UM2)
M1B-I: orthopyroxenite above M1B-H (subunit UM2)
M1B-J: orthopyroxenite above M1B-I (subunit UM2)
M1B-K: orthopyroxenite above M1B-J (subunit UM2)
M1B-L: orthopyroxenite above M1B-K (subunit UM2)
M1B-M: orthopyroxenite above M1B-L (subunit UM2)
M2EA: orthopyroxenite sample collected 15 m east of M1B samples (subunit UM2)

1999 at the base of the reverse-differentiated Basal Contact subunit. To date, S-bearing samples from the KO zone contain up to 5.4 g/t combined Pd+Pt+Au (Pd>>Pt>>Au), 2.3% Cu and 1.1% Ni (Peck et al., 1999). All samples analyzed to date contain disseminated sulphide mineralization and no massive sulphide mineralization has been assayed, although a 2 cm thick massive sulphide band occurs at the base of the MZ1 showing. Geochemical analyses obtained for samples from the recently discovered Leisure island showing (MZ12, discovered by M. Huminicki, L. Potter and G. Desharnais) contain up to 1.9 g/t combined Pd+Pt+Au, 2.1% Cu and 0.9% Ni (Table GS-9-1). At the time this report was written, PGE data were not available for most of the samples collected during the current program.

4) The available data indicate that the samples from the MZ12 showing have lower Pd/Pt ratios (average of 1.5 for five samples) than those from the MZ1 showing (generally >3). Also, most of the available samples from the KO zone are Cu-rich (average Cu:Ni ratio of approx. 2.1, although this ratio is quite variable, ranging from <0.1 to 8).
5) The large range in Cu:Ni ratios reflects variations in the chalcopyrite:pentlandite ratio.
6) The Cu and Ni concentrations generally decrease with increasing stratigraphic height (Fig. GS-9-5). This is in agreement with a simple density segregation process, whereby sulphide liquid droplets accumulate toward the base of a magma body.

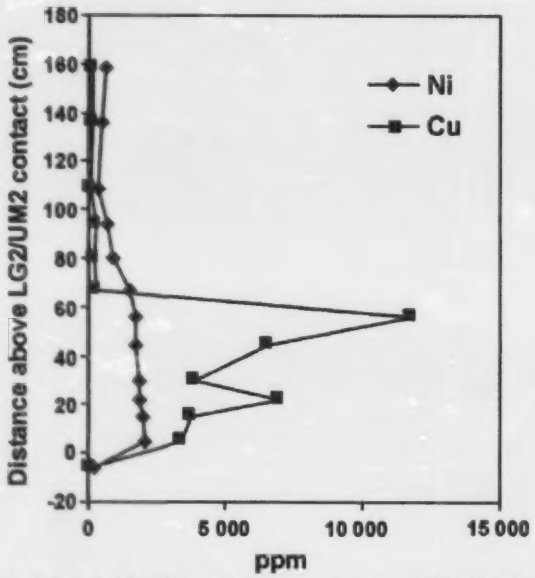

Figure GS-9-5: Variations in whole-rock Cu and Ni abundances with distance from the contact between the LG1 and UM2 subunits. Refer to Figure GS-9-3 for sample locations.

7) A plot of four major elements (oxides of Fe, Ca, Al and Mg) shows that Al and Fe vary similarly with stratigraphic height. In contrast, there is an antipathetic relationship between Al or Fe and Ca (Fig. GS-9-6). These trends are consistent with a pyroxene±plagioclase control on this part of the UM2, in contrast to the more typical olivine±pyroxene control evident in the overlying lherzolite sequence.
8) There is a very significant positive correlation ($r^2 = 0.97$) between Cu and Se contents in all KO zone samples collected from Leisure island. This suggests a strong sulphide control on the Cu content in this zone. The correlation between Ni and Se, although still significant, is weaker than that between Cu and Se. This may reflect a minor silicate control on Ni abundances in the KO zone (e.g. olivine).
9) Mantle-normalized multi-element plots for representative samples from the KO zone at Leisure island (Fig. GS-9-7) illustrate that most of the mineralized rocks have mantle-like trace-element compositions

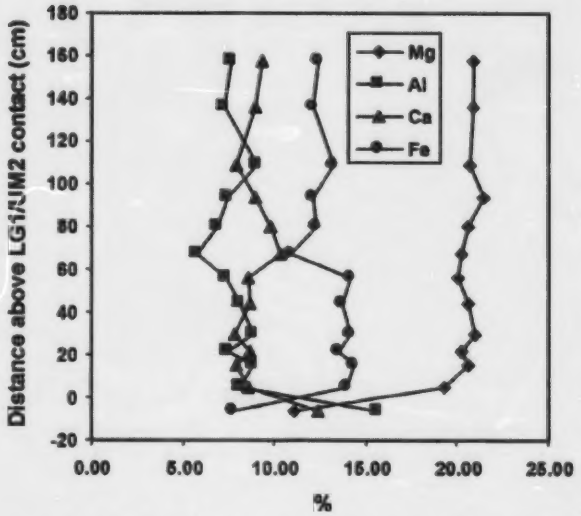

Figure GS-9-6: Variations in whole-rock MgO, CaO, Al_2O_3 and Fe_2O_3 contents with distance from the contact between the LG1 and UM2 subunits. Refer to Figure GS-9-3 for sample locations.

with prominent, negative Eu anomalies. These samples also display a positive Zr anomaly (Fig. GS-9-7). None of the samples analyzed to date show any significant enrichment in incompatible trace elements (e.g. La) that might be expected had the parent magmas assimilated significant amounts (bulk or partial melts) of country rock. This suggests that the magma was saturated in S prior to intrusion and that wall-rock assimilation may have been localized within the Basal Contact unit.

DISCUSSION: PRELIMINARY GENETIC MODEL FOR THE KO ZONE

A preliminary emplacement model is proposed that is generally consistent with both field observations and geochemical data:

1) Stage 1: Emplacement and crystallization of the reverse-differentiated Basal Contact unit, representing the first pulse of magma into the Fox River sill and effectively representing a broad chilled margin that locally grades into a petrologically complex magmatic breccia.
2) Stage 2: Emplacement of Cyclic unit 1, the first normally differentiated cyclic unit of the Marginal zone and the second major influx of magma into the proto-chamber of the Fox River sill. The Basal Contact unit and Cyclic unit 1 may, in fact, have formed from a single pulse of magma, but field relationships are presently inconclusive. Cyclic unit 2 magma differentiated through (?)flow-enhanced segregation of cumulus olivine to produce the observed ultramafic base (UM1) and leucogabbroic upper part (LG1). Late-stage leucotonalite liquid may have evolved at the temporary roof of the chamber or may reflect contact-related melting of the LG1 subunit during emplacement of Cyclic unit 2.
3) Stage 3: Sulphur-saturated ultramafic magma responsible for Cyclic unit 2 is injected between the anorthositic upper part of the LG2 subunit and the actual roof of the chamber (possibly underlain by leucotonalite from Cyclic unit 1), scouring the top of the UM1 subunit and creating local troughs within which dense S-liquid droplets are collected by flow-enhanced gravitational settling. The sulphide minerals concentrate along the lowermost part of the UM2 subunit, regardless of the presence of trough structures or late-stage mafic pods.
4) Stage 4: Conductive and/or convective heat loss across the solid anorthositic top of the LG1 layer helps advance crystallization of pyroxene in lieu of olivine. Early-formed pyroxene collects with the sulphide liquid in a mush zone at the base of the UM2 subunit. This

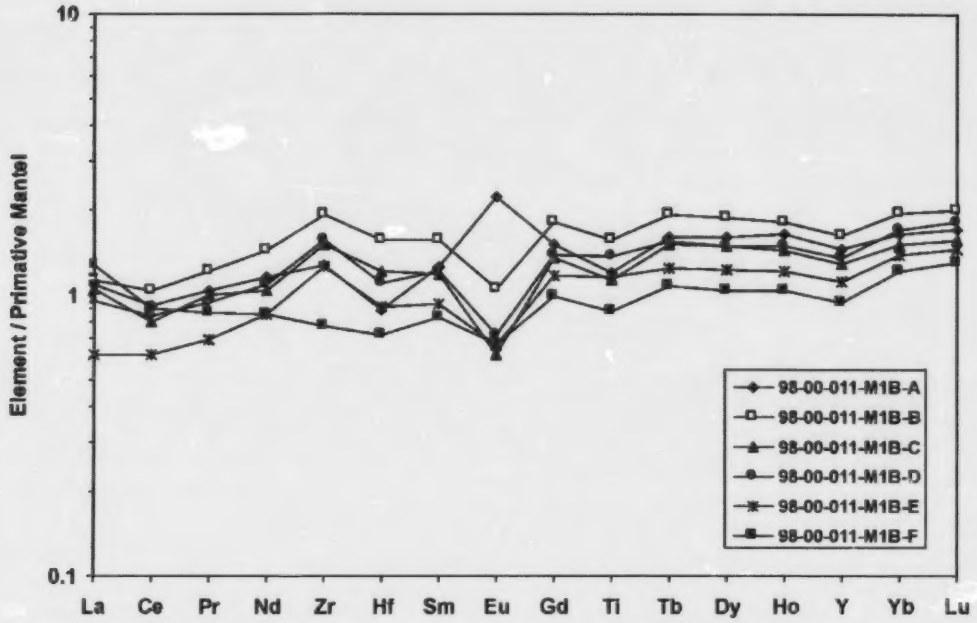

Figure GS-9-7: Condensed, mantle-normalized, multi-element plot for selected, representative samples from the contact between the LG1 and UM2 subunits. Refer to Figure GS-9-3 for sample locations. Sample 9800-011-M1B-A was collected from the northernmost part of the LG1 subunit. The remaining samples are from the lowermost part of the UM2 subunit.

heating may have promoted partial melting of the leucogabbroic rocks beneath the refractory, anorthositic top of the UM1 subunit, which then ascended across the LG1-UM2 contact and invaded the mushy base of the UM1 subunit. These pod-like injections entrained sulphide minerals and UM2 'mush', and crystallized as melagabbronorite-gabbronorite-leucogabbronorite-leucodiorite pods that have normal modal gradation and are locally mineralized (e.g. MZ1 pod). Several of these pods are variably textured, suggesting that the release of deuteric vapours from Cyclic unit 1 may have enhanced migration of the pod parent liquids upward into the UM2 subunit.

Ongoing research of the Marginal zone geology and mineralization is being carried out by the senior author (University of Manitoba). This research will provide new geological, mineralogical and geochemical observations that will allow the development of rigorous genetic models for the emplacement and mineralization of the KO zone. On a more regional scale, ongoing detailed investigations of the Fox River sill will continue to provide constraints on its emplacement and crystallization. One hypothesis currently being tested, which is consistent with many recent field and geochemical observations, is that the Fox River sill was emplaced on a subvertical, syn-rift, axial fault system and is therefore a major dyke.

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GS-10 CONTROLS ON GOLD MINERALIZATION ASSOCIATED WITH THE JOHNSON SHEAR ZONE AND AGASSIZ METALLOTECT, LYNN LAKE GREENSTONE BELT: SUMMARY OF CURRENT ACTIVITIES (PARTS OF NTS 64C/10, /11, /12, /14, /15, /16)

by C.J. Beaumont-Smith

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INTRODUCTION

The goal of this multiyear, multidisciplinary program is to improve our understanding of gold-mineralization processes in the Lynn Lake greenstone belt. By providing explorationists with metallogenic models for the formation of known gold deposits, it should be possible to narrow the focus of gold exploration in the Lynn Lake greenstone belt, thereby facilitating the discovery of new deposits.

The fieldwork, begun during the 1999 field season, focused on developing an understanding of the deformation history of the Johnson Shear Zone (JSZ) and the relationship between the shear zone and the associated gold mineralization. Fieldwork during the 2000 field season continued the 1999 work and began, through thesis projects, investigations of past-producing gold deposits associated with the JSZ and Agassiz Metallotect.

The projects undertaken in this program represent a collaborative effort involving researchers from the Manitoba Geological Survey (MGS; C.J. Beaumont-Smith, H.V. Zwanzig), the University of New Brunswick (D. Lentz and graduate student G. Ma), Laurentian University (B. Lafrance and graduate student R. Jones), the University of Manitoba (N. Halden and graduate student D. Rogge), the University of Windsor (A. Turek) and Black Hawk Mining, Inc.

CURRENT ACTIVITIES

Work during the 2000 field season focused on the tectonic evolution of the Lynn Lake greenstone belt and controls on gold mineralization. The following aspects of the geological and mineralization history of the belt are reported in this volume:

1) Structural and geological investigations of gold mineralization associated with the Agassiz Metallotect (Ma et al., GS-11, this volume; area 1, Fig. GS-10-1) are following up on earlier MGS efforts to understand the gold mineralization associated with the unique stratigraphy of the Agassiz Metallotect. Initial fieldwork focused on detailed structural analysis, geochemical sampling and diamond-drill core logging of the MacLellan gold deposit and Agassiz Metallotect stratigraphy west of the deposit.
2) Structural analysis of the Johnson Shear Zone in the Gemmell Lake–Dunphy Lakes area (Beaumont-Smith, GS-12, this volume; area 2, Fig. GS-10-1) is an ongoing regional study to document the western extension of the JSZ, including newly discovered gold mineralization within the shear zone.
3) Detailed structural analysis is being carried out on this area of newly discovered gold mineralization within the JSZ west of Gemmell Lake (Beaumont-Smith and Edwards, GS-13, this volume; area 3, Fig. GS-10-1).
4) Structural analysis and gold metallogeny of the Burnt Timber gold deposit (Jones et al., GS-14, this volume; area 4, Fig. GS-10-1) focuses on the structural and geochemical controls on gold mineralization in this deposit, which is hosted within the JSZ.
5) A structural and geochemical study of the Farley Lake gold deposit (Beaumont-Smith et al., GS-15, this volume; area 5, Fig. GS-10-1) examines a style of mineralization unique to the Agassiz Metallotect.
6) Gold and base-metal anomalies generated during a GIS analysis of compiled geophysical and geochemical data from the JSZ are being checked on the ground (Rogge et al., GS-16, this volume; areas 6, Fig. GS-10-1).
7) Geochemical and tectonic studies have been carried out in the Laurie Lake–Eager Lake area, on the boundary between the southwestern Lynn Lake greenstone belt and the Kisseynew Domain (Zwanzig, GS-17, this volume; area 7, Fig. GS-10-1).
8) A geochronological study focuses on providing high-precision U-Pb zircon age determinations (Turek et al., GS-18, this volume; areas 8, Fig. GS-10-1).

Investigations during the balance of this year will include high-precision geochemistry of samples collected during the field season, to determine the chemostratigraphy and characterize the alteration associated with the gold mineralization. Petrographic and microstructural analysis will provide information on the relationship between structural fabrics and gold mineralization, furthering the mesoscopic observations already made. Stable-isotope, micro-analytical and geochronological investigations are planned but have yet to be scheduled.

The age determinations reported in this volume represent significant additions to the geoscientific database for the Lynn Lake greenstone belt. It is hoped this work will be expanded in the coming years to address a significant gap in our understanding of the tectonic evolution of the greenstone belt. This has direct application to the ongoing mineral-deposit studies and future mineral exploration.

Initial efforts have significantly expanded the known extent of the Johnson Shear Zone, thereby increasing the amount of prospective auriferous stratigraphy. The identification of new areas of gold mineralization hosted by the western JSZ demonstrates the significant exploration potential of the Lynn Lake belt. The thesis projects and regional structural studies will be completed during the next three field seasons, and the program is scheduled for completion in 2003.

ACKNOWLEDGMENTS

The authors wish to thank the staff of Black Hawk Mining, Inc. for their technical assistance during the preparation and fieldwork portions of this program. In particular, the hands-on involvement of Paul Pawliw has greatly facilitated the various studies. Dan Ziehlke of Strider Resources Ltd.–Union Gold, Inc. is thanked for property access and many stimulating discussions. The thorough reviews of Herman Zwanzig and Eric Syme greatly improved the manuscript.

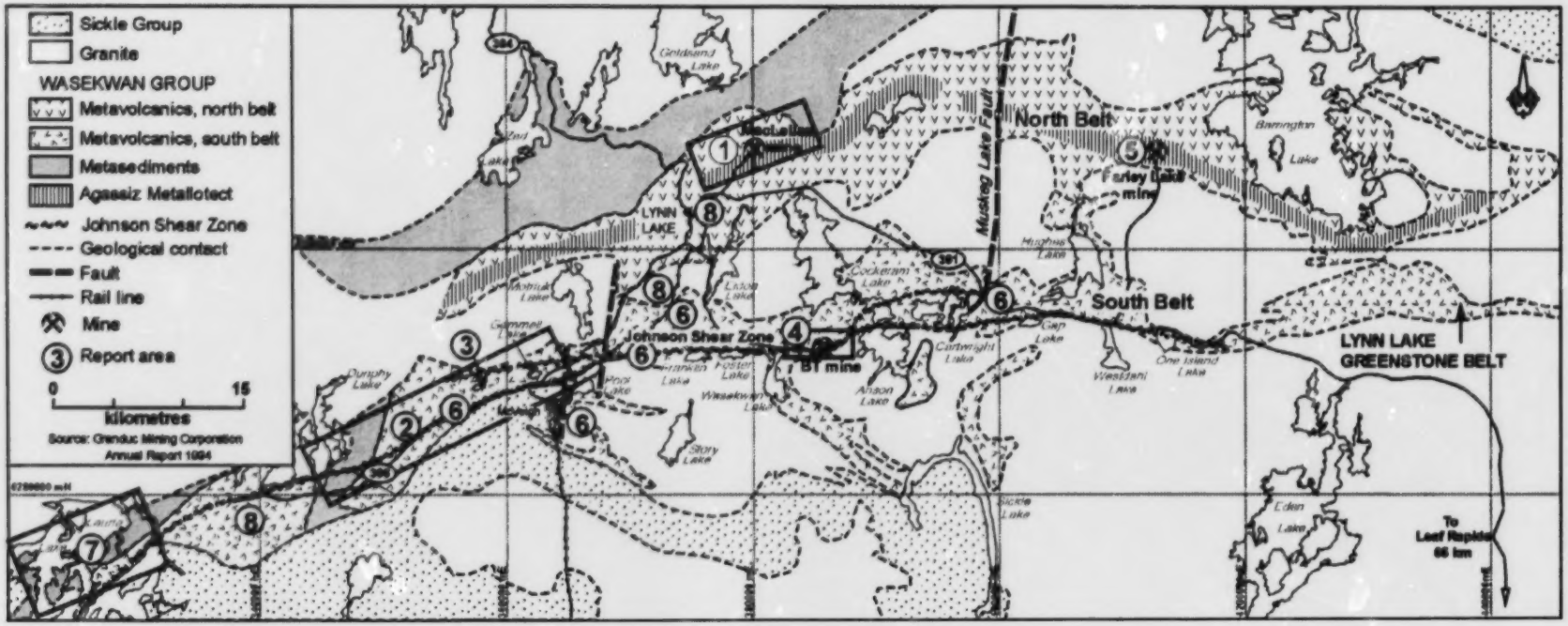

Figure GS-10-1: Geological setting of the Lynn Lake greenstone belt and the location of projects reported in this volume.

GS-11 PRELIMINARY STRUCTURAL ANALYSIS OF THE AGASSIZ METALLOTECT NEAR THE MACLELLAN AND DOT LAKE GOLD DEPOSITS, LYNN LAKE GREENSTONE BELT (PARTS OF NTS 64C/14, /15)

by G. Ma[1], C.J. Beaumont-Smith and D.R. Lentz[1]

Ma, G., Beaumont-Smith, C.J. and Lentz, D.R. 2000: Preliminary structural analysis of the Agassiz Metallotect near the MacLellan and Dot Lake gold deposits, Lynn Lake greenstone belt (parts of NTS 64C/14, /15); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 51-56.

SUMMARY

Preliminary field mapping and detailed structural analysis in the western portion of the Agassiz Metallotect have identified five generations of structural fabrics formed by discrete deformational events. Fabrics dominate the major layout of the rocks in the area. The area around the Dot and MacLellan gold deposits is dominated by tight to isoclinal, steeply east-plunging F_2 folds with a pronounced S_2 axial-plane foliation that strikes west-northwest and dips steeply to the north. The earlier S_1 fabric, which is now layer-parallel, is folded by these later folds; isoclinal F_1 folds are rare but locally preserved on the limbs of F_2 folds. The S_2 foliation was overprinted by S_3 fabrics, which are related to dextral kink folds. Open F_4 structures with a penetrative spaced cleavage are oriented perpendicular to the regional S_2 fabric. Late brittle deformation in the form of pseudotachylite breccia crosscuts all other deformational fabrics.

The stratabound gold mineralization occurs in both metavolcanic and metasedimentary rocks of the metallotect. The associated hydrothermal alteration is characterized by calcic amphibolitization, biotitization, muscovitization, silicification, carbonatization, chloritization and serpentinization of the host high-Mg-Ni-Cr basalt (picrite) and fine-grained sedimentary rocks. Gold mineralization seems to be principally associated with the S_2 fabric, as high assays are coincident with carbonate-quartz-sulphide veins crosscutting the sulphidic metasiltstone, and persistent low-grade gold mineralization also occurs in silicified picrite.

INTRODUCTION

The Agassiz Metallotect is an economically important zone of rocks that host significant gold deposits within the Paleoproterozoic Lynn Lake greenstone belt of northern Manitoba. This federal-provincial research project was initiated to complement and expand on earlier regional geological mapping, lithogeochemical sampling and gold deposit studies in the region. Due to our evolving understanding of the genesis of gold deposits in greenstone belts in general, there are presently conflicting genetic theories regarding the origin of the gold deposits, including those in the Lynn Lake region. Therefore, this mapping project in the northern part of the Lynn Lake greenstone belt was initiated to enhance understanding of its stratigraphic and structural evolution, with particular emphasis on the genetic relationship to gold mineralization in the region. This study also complements the detailed structural analysis begun in the southern part of the belt (Johnson Shear Zone; Beaumont-Smith, GS-10, -12, this volume; Beaumont-Smith and Edwards, GS-13, this volume; Jones et al., GS-14, this volume; Beaumont-Smith et al., GS-15, this volume).

The MacLellan and Dot Lake gold deposits are located about 7 km northeast of the town of Lynn Lake within the Paleoproterozoic Lynn Lake greenstone belt. Gold mineralization is hosted by a unique stratigraphic succession know as the Agassiz Metallotect (Fedikow and Gale, 1982), a subdivision of the metavolcanic and metavolcaniclastic Wasekwan Group. The Agassiz Metallotect has long been recognized as host to some important gold deposits, including MacLellan, Dot Lake, Farley and a number of other smaller occurrences (Fedikow, 1986). It has been the target of gold exploration since the 1950s and the focus of research on its geology, geochemistry and the relationship between these and gold mineralization (Fox and Johnson, 1981; Fedikow and Gale, 1982; Fedikow, 1983, 1986, 1992; Fedikow et al., 1986, 1991; Samson and Gagnon, 1995; Ferreira and Baldwin, 1997; Samson et al., 1999). However, the relationship between gold mineralization and deformation is poorly understood. This work is part of a program of detailed structural analysis of the Lynn Lake greenstone belt and the metallogeny of the gold deposits in the area. The purpose is to provide insight into the process involved in the formation of gold mineralization in the Lynn Lake area and provide new information to guide further exploration for gold deposits. Fieldwork during the 2000 field season included detailed 1:5000 geological mapping near the MacLellan and Dot Lake gold deposits, geochemical sampling, and the logging and sampling of a dozen diamond-drill holes from the area.

GEOLOGICAL SETTING AND STRATIGRAPHY

The Lynn Lake greenstone belt is subdivided into northern and southern supracrustal belts, with intervening intrusive complexes. The supracrustal rocks are composed mainly of Paleoproterozoic metamorphosed volcanic, volcaniclastic and sedimentary rocks that make up the Wasekwan Group (Bateman, 1945). These rocks have been intruded by plutonic rocks of the Pool Lake suite (Gilbert et al., 1980). The Agassiz Metallotect is included in Division D of the Wasekwan Group (Fedikow, 1986) and is characterized by a unique and well defined lithological assemblage, although facies changes are apparent (Fedikow, 1986). The metallotect is generally defined by an interlayered (intercalated) sequence of high-Mg-Ni-Cr basalt (picrite), sulphidic siltstone and iron-formation. This stratigraphic sequence has a strike length of over 65 km, and hosts the MacLellan and Dot Lake gold deposits in the western portion of the metallotect and the Farley Lake gold deposit in the eastern portion (Fedikow, 1983; Fedikow et al., 1986, 1991). In the vicinity of the MacLellan and the Dot Lake gold deposits, the stratigraphy comprises three units: hanging-wall mafic volcanic and volcaniclastic rocks, Agassiz Metallotect, and footwall mafic volcanic and volcaniclastic rocks (Fig. GS-11-1). The hanging-wall and footwall mafic rocks are texturally and geochemically very similar (Fedikow, 1986). Metamorphism within the northern part of the Lynn Lake greenstone belt is characterized by lower to middle amphibolite facies assemblages. The peak of metamorphism postdates the main deformation events and emplacement of the mineralization and associated alteration.

The hanging-wall mafic volcanic sequence comprises amygdaloidal and aphyric to plagioclase porphyritic basalt, autoclastic and heterolithic breccia, and thin mafic epiclastic units. The abundant mafic breccia within the hanging-wall sequence consists of matrix-supported clasts of dark green plagioclase porphyritic basalt, amygdaloidal basalt, aphyric basalt and quartz porphyritic rhyolite in a matrix of variable composition. The basaltic clasts (<40 cm) are rounded, elongated and locally strongly epidotized, whereas the rhyolite clasts are usually rounded and slightly elongated (Fig. GS-11-2). The breccia matrix ranges from fine-grained hornblende and plagioclase, very similar in composition and appearance to the clasts, to a very fine grained, silty matrix. The local preservation of primary depositional features, most notably bedding, together with the heterolithic nature of the breccia in the immediate MacLellan deposit area, suggests that this unit probably

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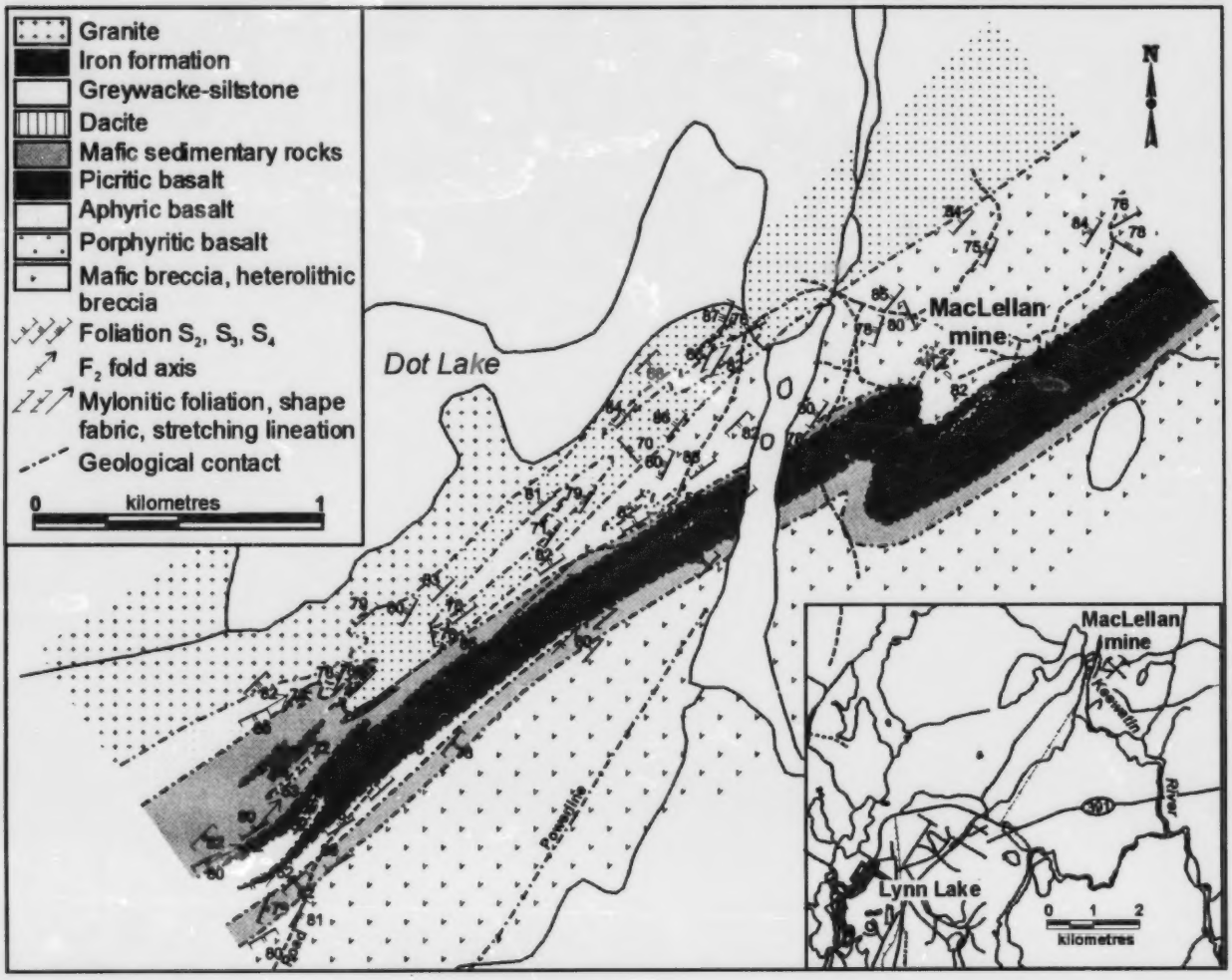

Figure GS-11-1: Geological map of the Dot Lake and MacLellan gold deposits area.

Figure GS-11-2: Basaltic epiclastic layer within heterolithic mafic breccia. S_2, which is defined by the stretched clasts, is crosscut by refracting S_4.

represents a debris flow. The coarse breccia north of the MacLellan headframe grades laterally (west) into finer grained breccia and finely laminated mafic epiclastic sedimentary rocks, consistent with more distal facies of this volcaniclastic unit.

This lateral facies change is mirrored by a change in the nature of the basalt in the stratigraphy. The hanging-wall porphyritic basalt in the MacLellan deposit area contains plagioclase phenocrysts, about 3 mm in size, that typically constitute approximately 20% of the rock. To the west of the deposit, the basalt becomes increasingly aphyric as the proportion of breccia decreases. To the north of the deposit, the phenocrysts are less abundant and smaller, with the basalt eventually becoming aphyric in the northern part of the study area.

The Agassiz Metallotect in the MacLellan–Dot Lake area consists of high Mg-Ni-Cr basalt (picrite), dark green (mafic) sedimentary rocks, greywacke, and minor dacite and oxide-facies iron-formation. The Agassiz Metallotect stratigraphy in the MacLellan–Dot Lake area is dominated by picrite in the immediate MacLellan mine area and by sedimentary rocks to the west.

The picrite has a highly variable mineralogical and textural nature, reflecting alteration and deformational changes throughout the study area. Picrite is recognized by its bright forest-green colour and very soft outcrop surface. Unaltered picrite is composed of coarse-grained actinolite (up to 5 mm) and lesser dark green hornblende in a fine- to coarse-grained chloritic groundmass. The picrite is generally strongly foliated, with the main foliation defined by the preferred orientation of amphibole. Rarely preserved pillows with well defined selvages and porphyritic cores indicate a submarine volcanic-flow setting (Fig. GS-11-3).

Mineralized and altered picrite is characterized by increasing degrees of biotitization, silicification and carbonatization. The mineralized zones exposed at the MacLellan and Dot Lake deposits are very highly strained, making an assessment of the nature of alteration and mineralization difficult. These zones generally comprise intercalated, lens-shaped domains of coarse-grained actinolite-chlorite (unaltered picrite), biotite-actinolite-talc-quartz and biotite-actinolite/hornblende-plagioclase-quartz with minor amounts of finely disseminated sulphide minerals (dominantly pyrite; cf. Gagnon, 1991). It is uncertain at this stage of the investigation whether the latter two constituents represent boudinaged and transposed, originally interbedded stratigraphic units (now altered) or transposed zones of varying degrees of alteration. The altered, biotite-rich domains are significantly more siliceous than the unaltered picrite, and contain a very penetrative fabric defined by the preferred orientation of biotite and amphibole.

Intercalated with the picrite are layers of well laminated biotitic greywacke, composed of quartz, plagioclase, biotite, and minor amphibole and disseminated sulphide minerals. Siltstone clasts are also present but are rare. Low-grade gold mineralization is associated with sulphide minerals, which locally approach 60% of the rock. Possible mafic sedimentary rocks associated with this unit have a bedded (laminated) texture and are dominated by metamorphic fine-grained amphibole and plagioclase, although these units may represent highly deformed mafic volcanic rocks.

Oxide-facies iron-formation, the second diagnostic rock type characterizing the Agassiz Metallotect, occurs as 10 to 30 cm thick units of magnetite-chert exhalative sedimentary rocks locally interbedded with the clastic sedimentary sequence. The finely laminated magnetite and chert layers clearly show the original bedding (Fig. GS-11-4). The competent chert layers were folded and boudinaged into discontinuous layers during deformation.

Minor units of grey to white dacite are found intermittently throughout the metallotect stratigraphy. These units are composed of fine-grained quartz and plagioclase. The nature of these units is not well understood and the contact relationships with surrounding rock types are unclear. These rocks locally appear conformable with sharp contacts, whereas, in other areas, they are characterized by diffuse contacts that appear discordant. The uncertainty regarding this unit revolves around the possibility that it may represent either felsic dykes with local alteration of their host, or transposed silicic alteration.

The footwall mafic volcanic rocks only outcrop in the southwest corner of the study area. They consist of fragmental basalt very similar in appearance to the hanging-wall sequence. The rock has a clastic texture, although it is typically massive in appearance with minor bedding features preserved. The porphyritic phases contain angular and unsorted plagioclase phenocrysts. There are cases where this fragmental basalt unit contains some smaller clasts of aphyric or porphyritic basalt.

STRUCTURAL ANALYSIS

Detailed structural analysis was conducted around the MacLellan and Dot Lake gold deposits in order to understand the postdepositional deformational history and its role in controlling the gold mineralization. Detailed mapping has identified five generations of deformational fabrics.

The preservation of early S_1 fabrics is rare due to the intensity of S_2. In siltstone, S_1 was observed as relict isoclinal folds on the limbs of F_2 folds, with its axial plane folded by F_2 folds. The F_2 event folds an older bedding-parallel tectonic fabric (S_1), which is locally preserved in S_2 crenulation cleavage microlithons.

The most penetrative fabric developed in rocks of the Agassiz

Figure GS-11-3: Pillowed porphyritic picrite. The pillow cores contain coarse-grained amphibole (after pyroxene), whereas the margin is composed of fine-grained chlorite.

Figure GS-11-4: Banded chert-magnetite iron-formation. Note boudinage of more competent chert layers. Back-rotation of chert boudins indicates dextral transcurrent shearing.

Metallotect is S_2, which generally forms a finely spaced to differentiated, steeply northwest-dipping foliation in the picrite and sedimentary rocks. It is defined in the coarse volcaniclastic rocks as a preferred orientation of the flattened clasts. The lack of S_1 preservation results from its widespread transposition to form an S_2 composite fabric. Zones of S_2 crenulation cleavage are generally restricted to F_2 hinge areas, reflecting the lower strains associated with the fold hinges. The S_2 fabric is axial planar to tight to isoclinal F_2 folds that plunge moderately to steeply to the east-northeast (Fig. GS-11-5). The F_2 folds represent the primary control on the distribution of units within the metallotect. The overall geometry of these folds is Z-asymmetrical.

The possible deformational control on the mineralization and alteration of the gold deposits hosted by the metallotect is suggested by the close association between distribution and timing of the introduction of mineralization and the development of D_2 high-strain (shear zone) fabrics. This association is evident at both the MacLellan and Dot Lake (K zones) deposits. In both cases, fabric development is characterized by the development of pre–peak metamorphic dextral shear fabrics, including dextral transcurrent S-C fabrics and shear-band cleavages (Fig. GS-11-6). Where the picrite is strongly altered, quartz and carbonate ribbons and veins define S_2, producing ribbon mylonite (phyllonite). Silicification and carbonatization of the picrite begins as syn-shear veins that become progressively transposed; this is a cyclical process that results in the wholesale silicification and carbonatization of the host. The transcurrent shear fabrics are accompanied by the development of steeply plunging to down-dip stretching lineation. These kinematics are consistent with those characterizing the Johnson Shear Zone (Beaumont-Smith and Rogge, 1999; Beaumont-Smith, GS-12, this volume), further supporting the possibility of the regional development of D_2 shear zones elsewhere in the Lynn Lake greenstone belt.

The field observations suggesting a link between D_2 fabric development and mineralization are reinforced by examination of drill core. Logging and sampling of a dozen diamond-drill holes shows that the mineralization is controlled by both lithology and hydrothermal alteration related to D_2 shear-related deformation. The higher average gold mineralization is associated with sulphidic greywacke and silicified picrite. No mineralization has been documented in the hanging-wall or footwall mafic volcanic rocks. Almost all the picrite-hosted gold mineralization is associated with quartz-carbonate-sulphide veins and the associated alteration. In the Dot Lake K_2 zone, gold mineralization is closely related to a 4 m wide zone of quartz-sulphide veining accompanied by silicification and intense biotitization, which is subparallel to S_2. The mineralized outcrops also show that the quartz-carbonate-sulphide veinlets and the alteration are subparallel to S_2. Although there are several later generations of quartz, quartz-carbonate and quartz-carbonate-chlorite veins, they do not seem to have sulphide or gold mineralization associated with them. At this point, based on these field-scale observations, it seems that gold mineralization is associated with hydrothermal

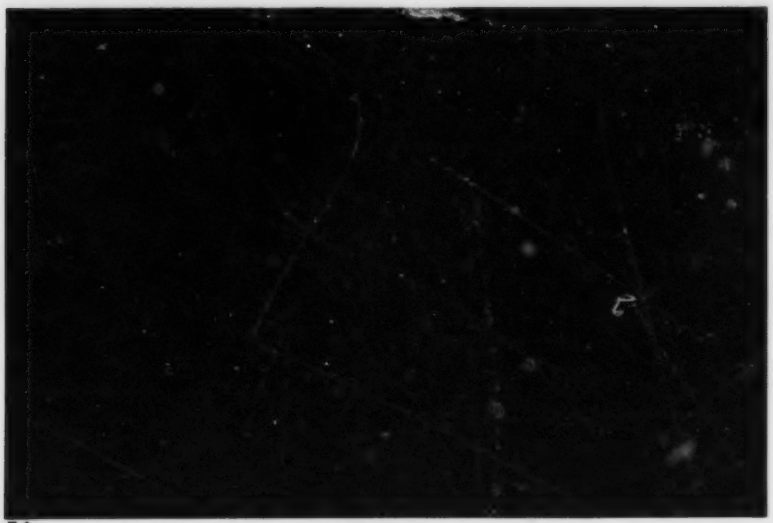

Figure GS-11-5: Z-asymmetrical F_2 folds in siltstone.

Figure GS-11-6: D_2 dextral S-C fabric and shear-band foliation in picrite.

activity related to late S_2 fabric development and shearing (*see also* Samson et al., 1999). However, it is still uncertain if this zone contained any pre-existing gold mineralization (epithermal style), as was postulated in earlier models (cf. Fedikow, 1986). If this was the case, remobilization of sulphide minerals and gold from pre-existing mineralization by focused fluid flow along this deformation zone may enhance earlier protore mineralization; however, further work is required to ascertain the most appropriate genetic model and discriminate between the two dominant hypotheses. Therefore, microstructural analysis will be done on the outcrop and drill core samples taken this year around the deposits, so as to examine the gold and base-metal mineralization and its relationship to deformation fabrics.

The D_2 fabric elements and the gold mineralization and alteration are overprinted by northeast-trending F_3 folds and axial-planar S_3 crenulation cleavage (Fig. GS-11-7). The F_3 folds are open to close in style and are uniformly Z-asymmetrical. They plunge moderately to the northeast and the associated northeast-trending S_3 crenulation cleavage dips steeply northwest. The crenulation microfolds generally have a chevron geometry, as do the F_3 folds in areas of elevated D_3 strain. Locally, narrow pseudotachylite zones follow S_3, but the exact timing of pseudotachylite development is less clear and may represent exploitation of the S_3 foliation as a pre-existing channelway, as opposed to being the result of D_3 strain.

Development of the D_3 fabric is broadly coincident with the peak of metamorphism. Metamorphic recrystallization overprints D_2 fabrics, whereas D_3 fabric development appears to span the growth of peak metamorphic assemblages (cf. Beaumont-Smith and Rogge, 1999). In picrite, actinolite and biotite porphyroblasts are randomly oriented in some rocks and define S_3 elsewhere, suggesting that the time span of porphyroblast growth reflects variations in local bulk chemistry.

Two post–peak metamorphic deformation events have been identified. A D_4 deformation produced north-trending, open F_4 folds and a weak S_4 foliation comprising crenulation cleavage, fracture cleavage and kink bands. The S_4 foliation commonly forms a conjugate set of kink bands or fracture cleavage. The final deformation produced predominantly east-trending brittle structures. These late structures generally take the form of zones of cataclasite, highlighted by the development of pseudotachylite, which form along the shear structure and are injected into the other intersecting brittle structures (Fig. GS-11-8). The cataclasite zones strike subparallel to the S_2 trend but dip slightly less steeply. The North shear that truncates the MacLellan main zone appears to be a D_5 structure, according to the descriptions of Gagnon (1991) and Samson and Gagnon (1995).

CONCLUSIONS

Preliminary structural analysis of the Agassiz Metallotect has identified

Figure GS-11-7: S_3 crenulation cleavage overprinting S_2 composite foliation in picrite.

Figure GS-11-8: D_5 pseudotachylite (arrow) developed in mylonitic sediments. Note the abundant shear bands in the finely laminated upper portion of the photo.

five generations of ductile and brittle-ductile fabrics. Near the Dot Lake and MacLellan gold deposits, D_2 fabrics are the dominant fabric elements and exercise primary control on the distribution of the stratigraphic units. The deformation associated with D_2 is locally intense and characterized by very high strains, as evidenced by the development of tight to isoclinal folds, the transposition of earlier S_1 foliations producing a composite S_2 foliation, and the local development of S_2 parallel dextral shear zones. It is these zones that are empirically most intimately associated with gold and base-metal mineralization and related hydrothermal alteration. Peak metamorphism is broadly coeval with F_3 folding, based on the presence of amphibole (±plagioclase±phyllosilicate) porphyroblasts along the associated fabrics. The late F_4 open folds and D_5 brittle fabrics do not appear to represent significant deformations in terms of the distribution of units or involvement in the gold-mineralizing event.

As an extension of this work, detailed petrology is planned to complement the geochemical analyses that are presently being obtained. About 200 samples were collected from available drill core and outcrops for polished thin section work, X-ray diffraction analysis, and geochemical analysis. This lab work should provide more clues to the genesis of the MacLellan and Dot Lake gold deposits.

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GS-12 STRUCTURAL ANALYSIS OF THE JOHNSON SHEAR ZONE IN THE GEMMELL LAKE-DUNPHY LAKES AREA, LYNN LAKE GREENSTONE BELT (PARTS OF NTS 64C/11, /12)

by C.J. Beaumont-Smith

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SUMMARY

The Johnson Shear Zone represents a major east-trending, pre–peak metamorphism shear zone located in the southern Lynn Lake greenstone belt. The shear zone is characterized by the development of dextral transcurrent shear-sense indicators and steeply plunging stretching lineations consistent with a dextral transpression deformational regime. There is mounting evidence of a long history of movement along the shear zone, with the majority of observed shear fabrics representing a major period of D_2 reactivation of an older shear zone. The Johnson Shear Zone has been extended west of Gemmell Lake, its previously delineated western limit, to the Dunphy Lakes area, where it represents the eastern strike extension of the previously defined Dunphy Lakes–Todd Lake Fault System. The inclusion of the Dunphy Lakes–Todd Lake Fault System results in a total shear-zone strike length exceeding 100 km.

Gold mineralization west of Gemmell Lake occurs in several widely spaced localities, hosted either by sheared and altered felsic volcanic rocks or by sheared and altered mafic volcanic rocks. The felsic-hosted mineralization is characterized by the presence of very fine grained acicular arsenopyrite, locally accompanied by finely disseminated pyrite, within zones of syn-shear quartz veins, silicification and weak sericitization of the host dacitic pyroclastic rocks. This is a different and previously unrecognized style of mineralization, compared with the syn-shear, pyrite-associated, mafic volcanic–hosted mineralization associated with the Johnson Shear Zone. The abundance of thin, laterally continuous, felsic volcanic units within the mafic-dominated stratigraphy affected by the shear zone west of Gemmell Lake indicates significant areas of excellent gold-exploration potential. The second style of mineralization, informally termed 'BT-style' (after the Burnt Timber gold deposit), is associated with finely disseminated pyrite in highly silicified and carbonatized, sheared mafic volcanic rocks. The silicification is accompanied by moderate biotitization, reflecting syn-shear potassic metasomatism.

INTRODUCTION

A program of structural analysis of the Johnson Shear Zone (JSZ) was initiated in 1999 with the goal of understanding better the deformational history of the shear zone and associated gold mineralization. Initial efforts focused on the central portion of the JSZ, an area known to host numerous gold deposits and showings. Structural analysis of the JSZ during the 2000 field season focused on delineating the shear zone beyond the western extent identified in 1999. These efforts included detailed analysis of newly discovered, shear-hosted gold mineralization west of Gemmell Lake.

The JSZ has been a focus of gold exploration since the 1940s and hosts a number of gold deposits and showings (Fig. GS-12-1), including the past-producing Burnt Timber (BT) deposit (Bateman, 1945; Milligan, 1960; Gilbert et al., 1980; Kenaley, 1982; Peck, 1986; Fedikow et al., 1991; Sherman, 1992; Richardson and Ostry, 1996; Peck and Eastwood, 1997; Peck et al., 1998). Confirmation of the westward continuation of the JSZ west of Gemmell Lake significantly increases the amount of prospective auriferous stratigraphy and, as the results of the first two field seasons demonstrate, there is excellent potential for the discovery of additional shear-hosted gold mineralization.

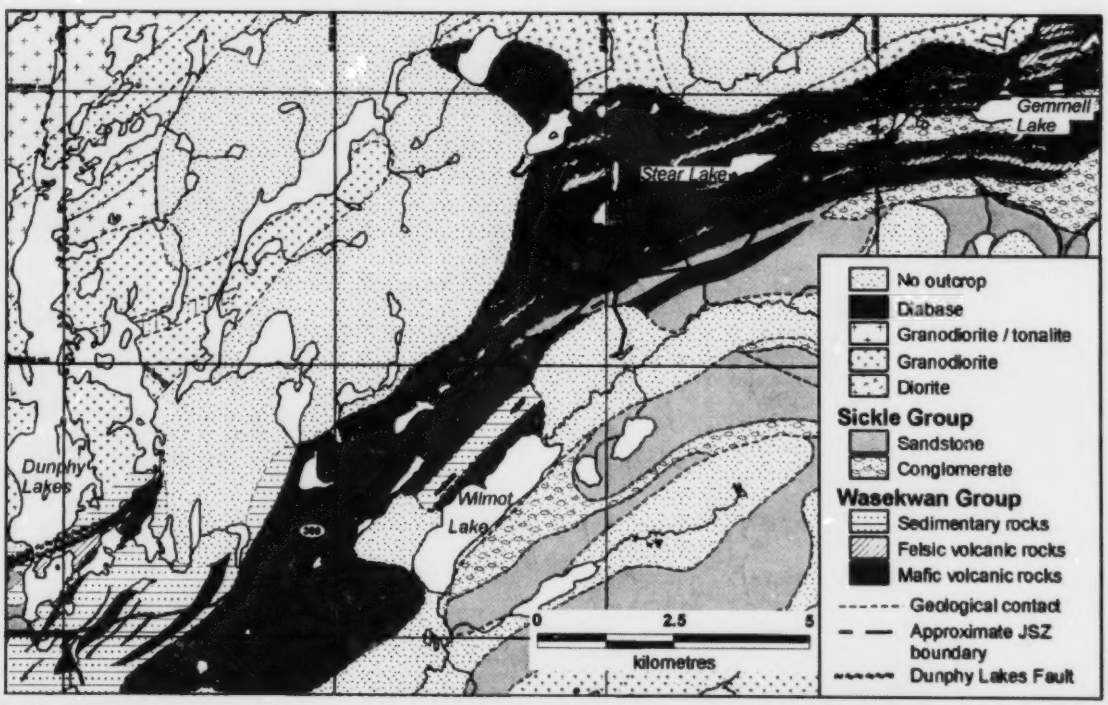


Figure GS-12-1: General geology of the Lynn Lake greenstone belt and location of the Johnson Shear Zone (after Gilbert et al., 1980).

GEOLOGICAL SETTING

The east-trending Johnson Shear Zone is located within the southern Paleoproterozoic Lynn Lake greenstone belt. It forms the boundary between the 1910 Ma (Baldwin et al., 1987) metavolcanic and metavolcaniclastic Wasekwan Group (Bateman, 1945) and the 1876 Ma (Baldwin et al., 1987) Pool Lake plutonic suite (Gilbert et al., 1980) in the central and eastern portions of the greenstone belt. The JSZ is manifest as an approximately 300 m wide zone of intense crenulation-cleavage development with an anastomosing mylonitic core. The localization of regional shear strain along the supracrustal-plutonic contact east of Gemmell Lake probably reflects a primary rheological control leading to the development of the JSZ. The absence of plutonic rocks along the southern margin of the southern greenstone belt west of Gemmell Lake removes the time-stratigraphic control that proved very useful in delineating the JSZ in the eastern portion of the greenstone belt.

West of Gemmell Lake, the JSZ is developed within the greenstone belt, affecting the supracrustal rocks, the unconformably overlying fluvial-alluvial arenaceous rocks of the Sickle Group (Norman, 1933; Gilbert et al., 1980), and numerous small plutonic bodies of undetermined age that intrude the Wasekwan Group. Age constraints on the development of the JSZ are limited, but shear-zone deformation is younger than the estimated age of the Sickle Group (1850–1540 Ma; Ansdell et al., 1999).

The metamorphic grade in the Lynn Lake greenstone belt ranges from upper greenschist facies in the eastern extent of the belt to upper amphibolite facies throughout the western portion of the belt. Peak metamorphism was attained relatively late in the deformational history, postdating the assembly of the greenstone belt and the development of the JSZ (Beaumont-Smith and Rogge, 1999). West of Gemmell Lake, the delineation of the JSZ is complicated by recrystallization associated with a middle to upper amphibolite facies metamorphic overprint that has removed many of the characteristic shear fabrics used to identify the JSZ at lower metamorphic grade.

The Wasekwan stratigraphy west of Gemmell Lake comprises a thick sequence of basaltic to andesitic volcanic and volcaniclastic rocks, thin intercalated dacitic to rhyolitic volcaniclastic units and metasedimentary rocks, which together form the Wilmot Lake volcanic and sedimentary rocks subdivision of Gilbert et al. (1980). The mafic rocks consist of massive flows and autoclastic and polymictic breccias, intercalated with minor layered volcaniclastic rocks. Thin dacitic to rhyolitic volcanic units and minor epiclastic sedimentary rocks are also intercalated with the mafic rocks. The Wilmot Lake subdivision is separated from the dacitic to rhyolitic volcaniclastic rocks of the Snake Lake dacite and the volcanic rocks hosting the Fox Lake mine by the Fox Road turbidite sequence, a thick sequence of fine-grained greywacke and conglomeratic greywacke. This sequence is interpreted to occupy the core of a large northeast-plunging synform (Gilbert et al., 1980).

The JSZ also affects the Sickle Group, which unconformably overlies the Wasekwan Group west of Gemmell Lake and forms the southern boundary of the greenstone belt in this area. The Sickle Group comprises a diverse stratigraphy of quartzofeldspathic sandstone, pebbly sandstone, hornblende-bearing sandstone and pelite, and heterolithic conglomerate. The contact between the Wasekwan and Sickle Groups is exposed in several locations along the southern boundary of the greenstone belt. The regional deformation has significantly modified the unconformity, making an assessment of the exact nature of the contact impossible. The Sickle Group also occurs as inliers within the Wasekwan Group; in the case of the large east-trending inlier centred on Gemmell lake, the contact between the two groups has experienced lower regional strain and a basal conglomerate is preserved.

STRUCTURAL ANALYSIS

The JSZ east of Gemmell Lake is characterized by the development of a mylonitic core within an approximately 300 m wide zone of intense crenulation cleavage and tight chevron folding (Beaumont-Smith and Rogge, 1999). West of Gemmell Lake, the kinematic framework remains consistent, although the shear zone is developed within the greenstone belt and appears to bifurcate into two major splays. The JSZ overprinted the assembly of the diverse tectonostratigraphy of the Wasekwan Group (Zwanzig et al., 1999) and the formation of the east-trending, overturned, antiformal geometry occupied by the Wasekwan Group prior to the intrusion of the Pool Lake suite (Gilbert et al., 1980). This complex sequence of deformations constrains the JSZ as a second- or possibly third-generation deformational event, assuming the older deformations represent discrete events. Therefore, to avoid possible confusion, the D_2 designation of Beaumont-Smith and Rogge (1999) for the JSZ will be used, since the delineation of older deformations requires additional information and techniques.

Overprinting relationships developed along the margins of the JSZ and in lower strain domains within the shear zone demonstrate the presence of an older foliation (S_1). The tight geometry of F_2 folds has generally resulted in the transposition of S_1 on the F_2 limbs, but S_1 slaty cleavage is preserved within S_2 crenulation microlithons from F_2 fold hinges and has been observed in the form of a strong clast preferred orientation in Wasekwan Group volcaniclastic rocks and Sickle Group conglomerate and pebbly sandstone (Fig. GS-12-2). The presence of S_1 can also be inferred from the widespread development of mafic tectonite at middle to upper amphibolite facies. The fine, very regular spacing of

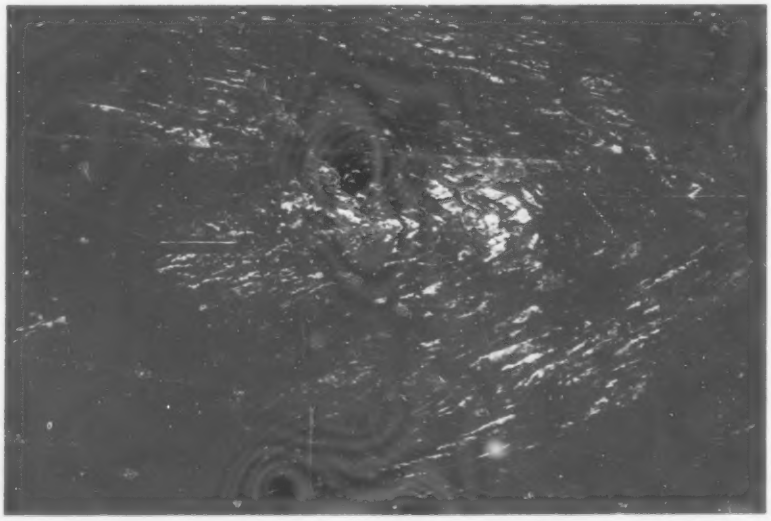


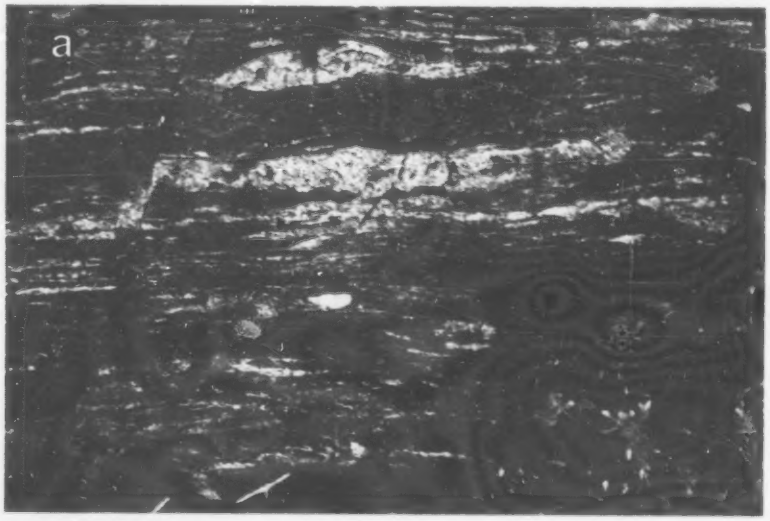
Figure GS-12-2: Overprinting of S_1 foliation, defined by clast preferred orientation, by tight F_2 fold on the margin of the shear zone. Note pen for scale.

alternating ferromagnesian and plagioclase-rich domains that make up the tectonites probably represents recrystallized crenulation cleavage.

The high-strain core of the JSZ is characterized by the development of anastomosing zones of mylonitic fabrics within a broader zone of intense crenulation-cleavage development. The enveloping zone of crenulation cleavage folds an older mylonitic fabric of uncertain age. This older fabric may represent mylonitic fabrics developed during earlier stages of D_2, reflecting the transient nature of shear zone development and the continual shifting (pulsation) of the high-strain portion of the shear zone. Alternately, the older fabrics may represent the products of an older (D_1) period of mylonitization. In the case of the latter, the intense crenulation and mylonitization associated with the JSZ represent the reactivation of a pre-existing shear zone of unknown age or kinematics. It is hoped that microstructural analysis will help resolve this problem.

An important characteristic of the JSZ is the development of dextral transcurrent shear-sense indicators on the horizontal surface and steeply plunging to down-dip stretching lineations (Fig. GS-12-3). The orientation of the stretching lineation on the S-surface and the ridge-in-groove striation on the C-plane (Lin and Williams, 1992) are generally subparallel. The stretching lineation developed in coarsely clastic rocks has a pronounced oblate geometry, indicating significant flattening strains. Fabrics observed parallel to the stretching lineation (X-Y plane of the finite strain ellipsoid) are generally symmetrical with the exception of shear bands, which dip more shallowly than the C-plane, suggesting a component of north over south dip-slip movement. The geometrical relationship between the shear-sense indicators and stretching lineation is generally interpreted to be the product of transpressional shear (Lin et al., 1998), a regime of shear-zone boundary-parallel transcurrent movement accompanied by a large amount of shear-zone–normal shortening.

The JSZ is overprinted by at least three discrete deformations. The most penetrative is D_3, which produced northeast–trending, open to close, Z-asymmetrical chevron folds and associated crenulation cleavages (Fig. GS-12-4). The D_4 deformation produced north-trending open folds and conjugate kink bands, and is most likely responsible for the large-scale open warps of the shear zone and the greenstone belt. The youngest structures overprinting the shear zone are D_5 pseudotachylite zones, which occur as thin (1–3 mm) shear-zone–subparallel seams or as breccia matrix within pseudotachylite-bound zones up to 30 cm wide within the shear zone (Fig. GS-12-5). The distribution of pseudotachylite has a strong spatial association with the JSZ, and it has not been identified outside the shear zone in the southern Lynn Lake belt. This association suggests that the development of pseudotachylite represents the late semibrittle reactivation of the shear zone. The sense of D_5 shear, as determined by the analysis of synthetic and antithetic Riedel shears within the pseudotachylite zones (Tchalenko, 1970), suggests predominantly sinistral strike-slip displacement of undetermined magnitude.



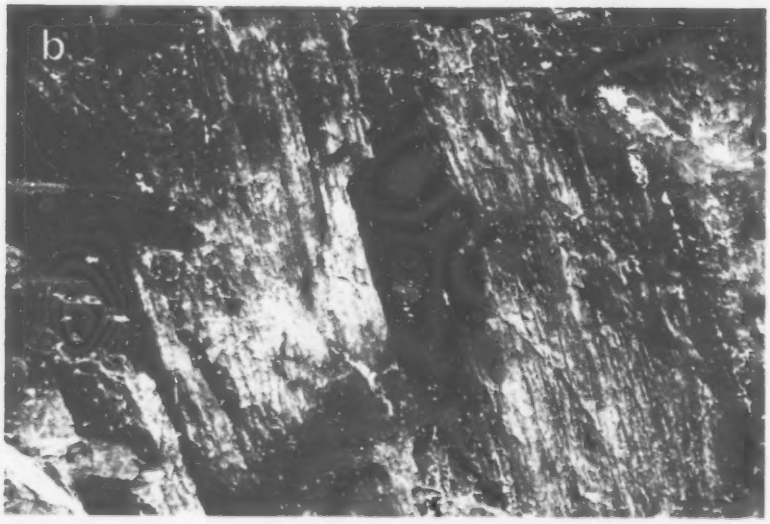


Figure GS-12-3: Kinematic characteristics of the Johnson Shear Zone include: a) dextral transcurrent shear-sense indicators such as asymmetrical boudinage, and b) steeply plunging stretching lineations.

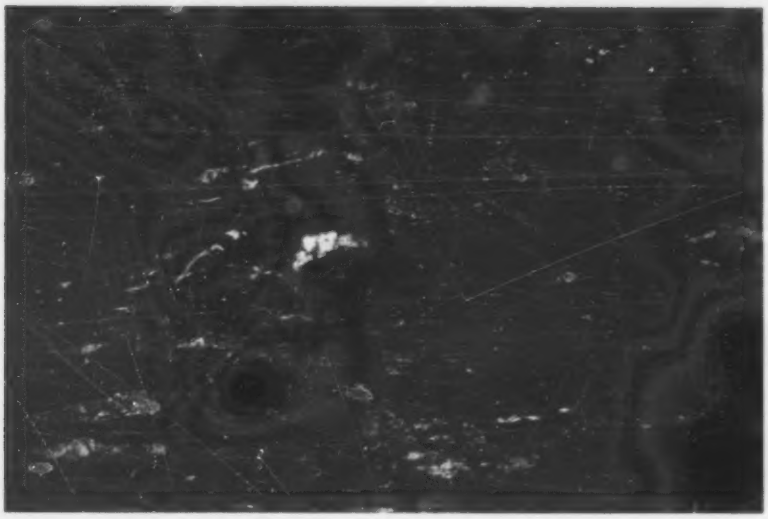


Figure GS-12-4: Close F_3 fold and axial-planar crenulation cleavage overprinting the S_2 mylonitic fabric.

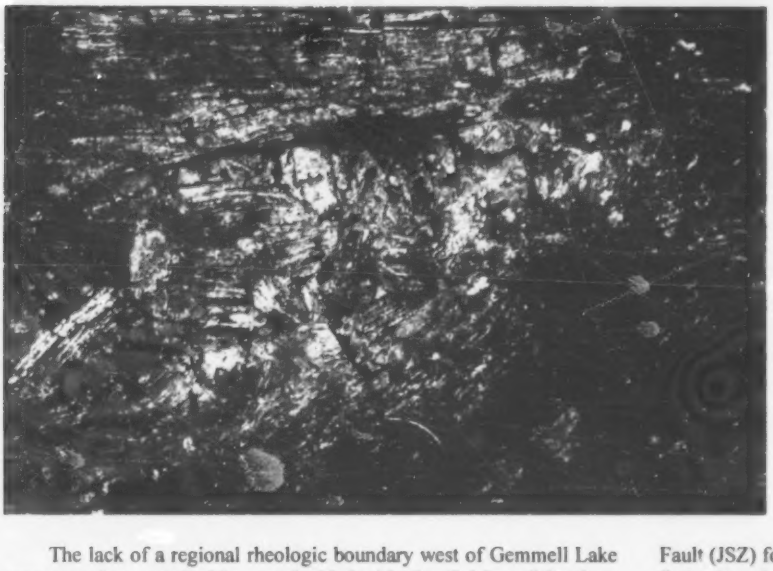


Figure GS-12-5: D_5 pseudotachylite breccia developed in Sickle Group feldspathic greywacke.

The lack of a regional rheologic boundary west of Gemmell Lake appears to have resulted in, or coincided with, the division of the shear zone into two major splays. One zone follows the south side of Gemmell Lake and proceeds west, south of Stear Lake. The delineation of the southern JSZ splay is facilitated by the presence of an inlier of Sickle Group arkosic and pebbly greywacke immediately north of Highway 396, between Stear Lake and Wilmot Lake. Originally interpreted as Wasekwan Group (Gilbert et al., 1980), additional outcrop exposed during recent forest fires exposes pebbly greywacke containing granodioritic clasts typical of the Sickle Group. This unit occupies the footwall to the JSZ and displays a marked south to north increasing strain gradient. The delineation of the JSZ west of Wilmot Lake is hampered by extensive glacial cover, but it is recognized again along the southeastern shore of east Dunphy Lakes. This westernmost exposure of the JSZ trends along the strike of the Dunphy Lakes Fault (Gilbert et al., 1980) and probably represents the eastern strike extension of the fault. The Dunphy Lakes Fault was inferred by Gilbert et al. (1980) to address stratigraphic juxtaposition east of the Fox Lake mine and has not been observed directly (H. Zwanzig, pers. comm., 2000).

The eastern strike extension of the Dunphy Lakes Fault, now the southern JSZ, is exposed along the northern limb of a macroscopic F_2 antiform cored by Snake Lake dacite (Gilbert et al., 1980). Detailed mapping of this area (also recently burned) confirms the fold interpretation originally advanced by Gilbert et al. (1980). The Dunphy Lakes Fault (JSZ) forms the contact between the dacite and the banded oxide-facies iron-formation and siliceous sedimentary rocks to the north and transects the dacite along the northern Z-asymmetrical limb of the fold. The shear zone is recognized by an approximately 20 m wide zone of intensely foliated and quartz-veined porphyritic dacite within a broad zone of strongly foliated dacite. The foliation development includes abundant asymmetrical pressure fringes around the phenocrysts, quartz-ribbon development in the matrix and a large amount of very fine quartz veins. The quartz veins are isoclinally folded, and asymmetrical boudinage of the veins and rare shear bands have been identified. Dextral transcurrent shear-sense indicators and steeply plunging to down-dip stretching lineations characterize the shear zone, consistent with the JSZ kinematic framework.

Interpretation of the Dunphy Lakes Fault as the western continuation of the JSZ has several significant stratigraphic and structural implications. The Dunphy Lakes Fault is interpreted to merge with or become the Todd Lake Fault in the Fox Lake mine area (Gilbert et al., 1980). Reconnaissance mapping of the Todd Lake Fault along the north shore of Todd Lake indicates that this structure maintains many of the characteristics of the JSZ. Most importantly, the Todd Lake Fault displays pre–peak metamorphic dextral transcurrent shear-sense indicators and steep stretching lineations consistent with the dextral transpressional kinematic framework that characterizes the JSZ. The Todd Lake Fault forms the structural hanging wall of the Fox Lake

volcanogenic massive sulphide (VMS) deposit and has been previously described as a largely ductile D_2 fault (Olson, 1986). The recognition of the Dunphy Lakes and Todd Lake faults as the western strike extension of the JSZ adds significantly to potential auriferous stratigraphy, and the understanding of the kinematics of the JSZ provides constraints on the search for additional VMS mineralization associated with the Fox Lake deposit, north of the JSZ.

The second JSZ splay forms a parallel zone of mylonitic fabrics that transects Gemmell Lake and continues west, north of the Sickle Group outlier that underlies Gemmell Lake (Fig. GS-12-1). This northern branch of the JSZ is poorly exposed but is recognized by the development of an approximately 100 m wide zone of intensely foliated felsic and mafic volcanic rocks with mylonitic fabrics that demonstrate the same kinematic framework as the southern branch (Fig. GS-12-6). The northern branch has been traced in scattered outcrop for a distance of approximately 5 km west of Gemmell Lake. Delineation of the northern splay farther to the west has been hampered by very limited outcrop, but it appears traceable by aeromagnetics where it passes through the southern end of Irene Lake.

GOLD MINERALIZATION

One of the most significant findings from this summer's fieldwork is the identification of auriferous stratigraphy within the JSZ west of Gemmell Lake. Stratabound gold mineralization is hosted by a 1 to 25 m thick felsic pyroclastic unit, intercalated with porphyritic gabbro, in the Gemmell Lake to Stear Lake portion of the JSZ (Beaumont-Smith and Edwards, GS-13, this volume). The felsic pyroclastic unit is dacitic in composition, locally garnetiferous and comprises various volcanic facies spanning a range from fine ash tuff to coarse lapilli tuff. Although tightly folded and highly sheared, the facies variation appears to be primarily lateral; vertical facies variation is minimal.

Gold mineralization is shear-hosted and associated with fine-grained acicular arsenopyrite ± finely disseminated pyrite in zones of silicified sheared dacite. The silicification is in the form of syn-shear quartz ribbons and quartz veins. The quartz ribbons form parallel to the mylonitic shear planes and, locally, the density of quartz ribbons produces bulk silicification. The quartz veins represent a spectrum of emplacement ages with respect to shear-zone development. The veins appear to nucleate as fine veinlets parallel to the S-surface, followed by their transposition parallel to the C-planes. Later stages of vein emplacement involve strongly asymmetrically boudinaged, coarsely polygonized quartz veins, with shear bands forming the terminations of the individual boudins (Fig. GS-12-7). Associated with the silicification is weak sericitic alteration, generally recognized by a beige weathered colour and increase in muscovite content (metamorphosed sericite). The weathered surface locally takes on a slight greenish hue, possibly reflecting a fuchsitic component to the alteration. Similar arsenopyrite-associated gold mineralization is reported in the east Dunphy Lakes

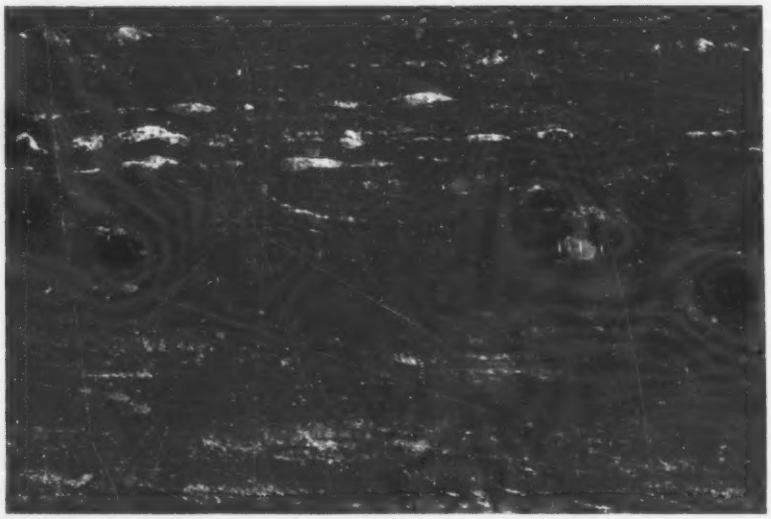


Figure GS-12-6: Dextral asymmetrical boudinage of quartz veins in felsic volcaniclastic rock affected by the northern splay of the Johnson Shear Zone.

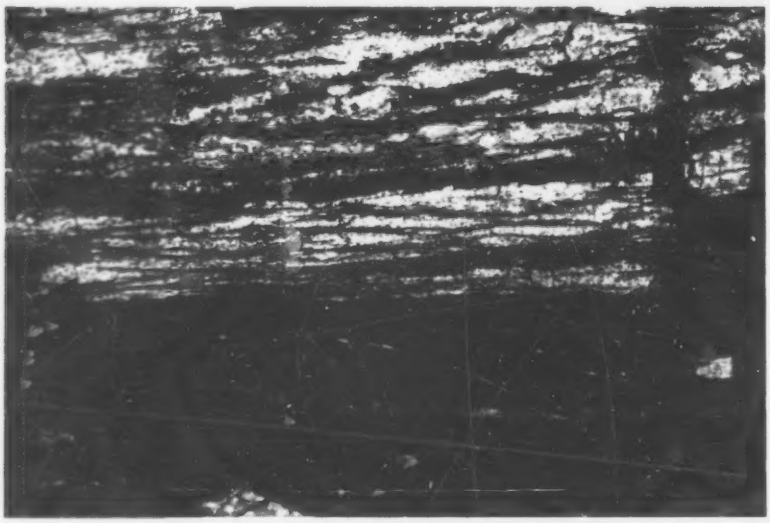


Figure GS-12-7: Dextral shear bands and asymmetrical boudinage of quartz veins in moderately sericitized and mineralized, dacitic, fine ash tuff.

portion of the southern JSZ, along the contact between mylonitic Snake Lake dacite and interbedded oxide-facies iron-formation and siliceous sedimentary rocks (Ferreira, 1993).

A second mineralized zone has been outlined north of Wilmot Lake, along Highway 396. The mineralization in this portion of the JSZ is associated with moderately to intensely Fe-carbonatized, silicified and sheared mafic volcanic rocks. The exact nature of the mineralization has yet to be determined, but appears to consist of finely disseminated pyrite within intensely silicified zones. The silicification and carbonatization occur as syn-shear veins, although the silicification has a spatial relationship with the emplacement of syn-shear feldspar-phyric felsic dykes and locally is sufficiently intense to produce bulk silicification. Silicified and sheared mafic volcanic host rocks have a pronounced purplish hue, suggesting the presence of very fine-grained biotite resulting from potassic metasomatism accompanying the silicification. The mineralization and alteration in this area are strikingly similar to those characterizing the BT deposit.

These widely spaced examples of shear-hosted gold mineralization associated with felsic and mafic volcanic rocks demonstrate the potential for the discovery of additional mineralization within the newly delineated western portion of the JSZ. The dacite pyroclastic unit west of Gemmell Lake occupies the same position within the southern JSZ as a similar unit immediately south of Gemmell Lake identified in 1999. This observation dictates a strike length for the zone exceeding 5 km in an area of discontinuous outcrop. The BT-style alteration, hosted by mafic volcanic rocks, can be traced for several kilometres along Highway 396 before leaving the road where it becomes lost under heavy glacial cover. The strike length of this alteration system is unknown.

CONCLUSIONS

The Johnson Shear Zone (JSZ) has been delineated west of Gemmell Lake and, with the incorporation of the Dunphy Lakes–Todd Lake fault system (based on their similar kinematic framework), this gives the JSZ a strike length of more than 100 km. The JSZ is a major D_2 structure characterized by the development of dextral transcurrent shear-sense indicators and steep stretching lineations. The kinematics of the JSZ are interpreted to be the result of dextral transpression. There is increasing evidence that the JSZ represents the D_2 reactivation of an older shear zone.

The western strike extension of the JSZ hosts several newly discovered zones of gold mineralization. Sheared mafic volcanic rocks in the Wilmot Lake area display alteration and mineralization similar in style to the BT deposit. A second, previously unrecognized mineralization style involves gold associated with finely disseminated acicular arsenopyrite and accompanied by syn-shear silicification and sericitization of dacitic volcaniclastic rocks. The discovery of gold mineralization along the western strike continuation of the JSZ west of Gemmell Lake demonstrates the exploration potential of the JSZ beyond its previous known extent.

ACKNOWLEDGMENTS

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GS-13

DETAILED STRUCTURAL ANALYSIS OF THE JOHNSON SHEAR ZONE IN THE WEST GEMMELL LAKE AREA (NTS 64C/11)

by C.J. Beaumont-Smith and C.D. Edwards[1]

Beaumont-Smith, C.J. and Edwards, C.D. 2000: Detailed structural analysis of the Johnson Shear Zone in the west Gemmell Lake area (NTS 64C/11); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 64-68.

SUMMARY

Gold mineralization hosted by the Johnson Shear Zone in the west Gemmell Lake area is associated with finely disseminated arsenopyrite in sericitized and silicified volcaniclastic dacite intercalated with porphyritic gabbro. The distribution of the dacite is controlled by tight to isoclinal F_2 folds, with an overall Z-asymmetry and moderate easterly plunge, that have produced considerable fold thickening of the mineralized horizon. The style of mineralization and alteration differs from previously documented styles associated with the Johnson Shear Zone and demonstrates the unrealized gold-exploration potential of the Lynn Lake greenstone belt.

INTRODUCTION

The Johnson Shear Zone (JSZ) is a major east-trending structure in the southern Lynn Lake greenstone belt (Fig. GS-13-1). The JSZ has been the focus of gold exploration since the 1940s (Bateman, 1945) and hosts the depleted Burnt Timber deposit (Richardson and Ostry, 1996). Earlier exploration for shear-hosted gold was concentrated on the then-known extent of the JSZ east of Gemmell Lake, where the shear zone forms the boundary between the greenstone belt and the Pool Lake suite plutonic rocks to the south (Gilbert et al., 1980). The recent delineation of the JSZ west of Gemmell Lake (Beaumont-Smith and Rogge, 1999; Beaumont-Smith, GS-12, this volume) has resulted in the discovery of shear-hosted gold mineralization in several new, widely separated locations.

Gold mineralization hosted within the JSZ in the west Gemmell Lake area represents a different style than that previously recognized in the Lynn Lake greenstone belt. Gold mineralization east of Gemmell Lake can be characterized by two end members: 1) gold associated with finely disseminated pyrite, accompanied by silicification, biotitization and carbonatization of the hosting, sheared, mafic volcanic rocks (BT style; Peck and Eastwood, 1997; Peck et al., 1998); and 2) gold associated with finely disseminated pyrite and lesser arsenopyrite, accompanied by silicification and variable carbonatization of the hosting, sheared, felsic dykes and intrusive rocks (Bonanza style; Peck, 1984, 1985, 1986).

The mineralization in the west Gemmell Lake area is associated with fine-grained acicular arsenopyrite and lesser pyrite accompanying the emplacement of syn-shear quartz veins and bulk silicification of weakly to moderately sericitized, dacitic volcaniclastic rocks. The objectives of this study are to 1) characterize this style of mineralization and alteration; 2) document the deformation history and structural geometry in the area of mineralization; and 3) determine the relationship, if any, between the deformation, alteration and mineralization.

GEOLOGICAL SETTING

The Paleoproterozoic Lynn Lake greenstone belt is an area of metamorphosed volcanic and sedimentary rocks that extends for 130 km from Laurie Lake in the west to Magrath Lake in the east. The belt has been subdivided into northern and southern, dominantly volcanic sequences of the Wasekwan Group (Bateman, 1945), separated by intrusive rocks of the Pool Lake suite (Gilbert et al., 1980). The Wasekwan Group was isoclinally folded into an east-trending anticlinorium prior to the intrusion of the Pool Lake suite (Gilbert et al., 1980). The Wasekwan Group and Pool Lake suite are unconformably overlain by an upper sedimentary succession of fluvial-alluvial conglomerate and sandstone that

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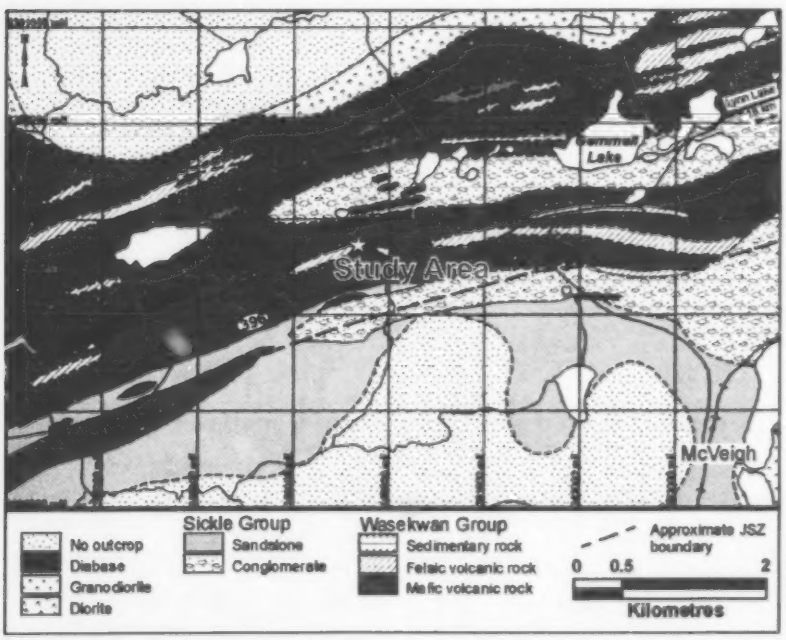

Figure GS-13-1: General geology of the Lynn Lake greenstone belt west of Gemmell Lake and location of the Johnson Shear Zone. The area of this study is highlighted (after Gilbert et al., 1980).

constitute the Sickle Group (Norman, 1933).

Metamorphic grade in the Lynn Lake belt ranges from upper greenschist to upper amphibolite facies. There is a general increase in metamorphic grade from east to west. The Gemmell Lake area underwent middle amphibolite facies peak metamorphism, following the development of the JSZ, that resulted in the recrystallization of primary and shear-zone fabrics. The timing of the metamorphic peak is broadly coincident with D_3 (Beaumont-Smith and Rogge, 1999).

Gold mineralization associated with the JSZ west of Gemmell Lake is hosted by sheared, dacitic, volcaniclastic rocks intercalated with variably deformed porphyritic gabbro within the southern Lynn Lake greenstone belt. The dacite is a fine-grained pyroclastic unit composed of fine to coarse ash with rare, thin, fine lapilli horizons. It has a white to beige weathered surface and is commonly strongly micaceous, the result of metamorphic recrystallization of sericitic alteration. The thickness of this unit varies from decimetres to approximately 25 m and is repeated several times across strike. Intercalated with the dacite is medium to coarsely porphyritic gabbro (now amphibolite). This unit has been interpreted as gabbro, based on the coarse grain size and local preservation of primary layering. These features are not unique to intrusive rocks, however, and this unit could possibly represent a mafic volcaniclastic rock. The gabbro is variably deformed, most likely the result of heterogeneous D_2 strain distribution and metamorphic recrystallization of the gabbro. Discrete high-strain zones are developed in the gabbro, producing fine-grained, finely laminated mafic tectonite units and narrow protomylonitic (porphyroclastic) zones.

STRUCTURAL ANALYSIS

Detailed structural analysis was conducted on an area of semicontinuous outcrop that includes several outcrops of newly discovered gold mineralization within the JSZ. The deformational history of the JSZ is characterized by the development of transcurrent, dextral shear-sense indicators and steep stretching lineations as a result of the concentration of regional D_2 deformation into an east-trending zone of very high strain (Beaumont-Smith and Rogge, 1999). Fabric development in the west Gemmell Lake area is consistent with that documented along the southern margin of the greenstone belt east of Gemmell Lake. The rocks underlying the study area are strongly foliated and lineated, and distribution of the units is controlled by tight Z-asymmetrical folds (Fig. GS-13-2).

The dacitic rocks are intensely foliated and mylonitic fabrics are commonly preserved. The main foliation, interpreted as S_2, is a finely spaced, muscovite-rich discrete shear foliation (C-plane) and quartz-ribbon orientation. Locally, dextral S-C fabrics are preserved in coarse ash layers, with the long dimension of the clasts defining the shape fabric. The high shear strain has resulted in the emplacement of a large number of fine quartz veinlets and ribbons at various times during D_2. Early syn-D_2 quartz veins are isoclinally folded and asymmetrically boudinaged, with D_2 dextral shear bands forming the boudin terminations (Fig. GS-13-3). Late-D_2 quartz veins have an anticlockwise orientation with respect to the S_2 mylonitic foliation and show the initial stages of asymmetrical boudinage. The intense D_2 deformation is characterized by steeply plunging to down-dip lineations. Lineations defined by stretched lapilli and quartz rodding plunge parallel to the striations on the C-planes (ridge-in-groove lineation).

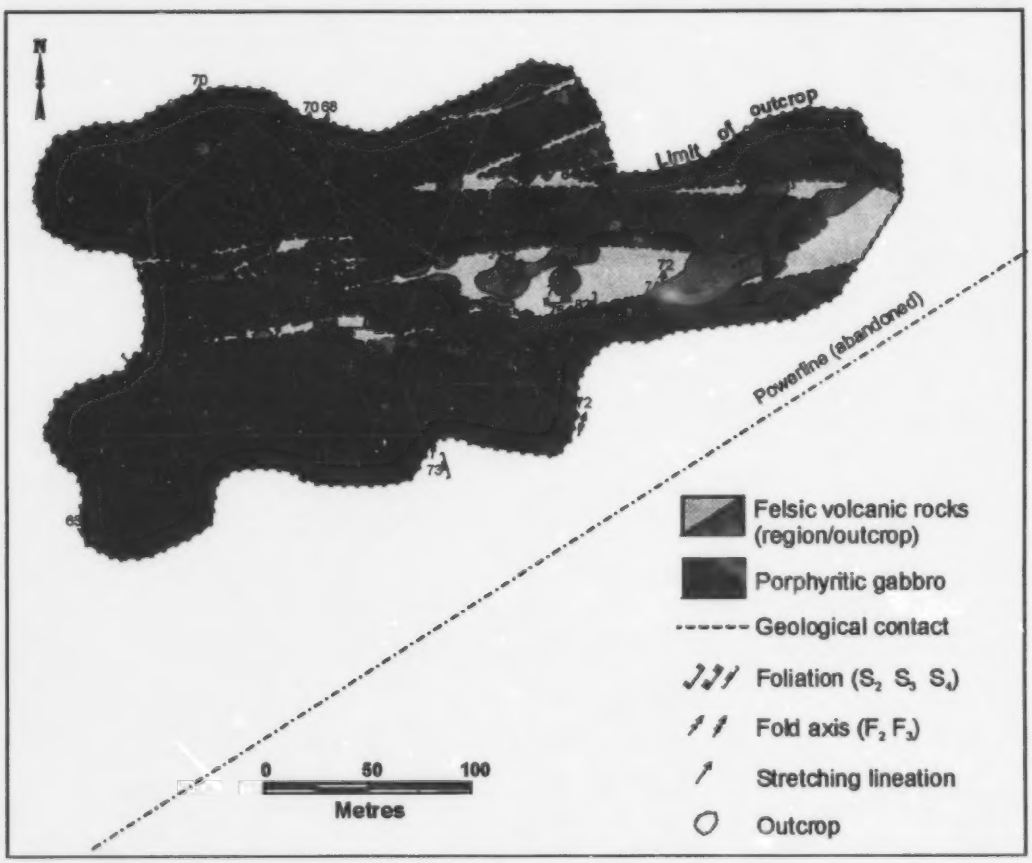

Figure GS-13-2: Detailed geology of the west Gemmell Lake area of newly identified gold mineralization.

Figure GS-13-3: Dextral shear bands and asymmetrical boudinage of quartz veins in host sericitized dacite.

The D_2 shear fabrics are overprinted by several generations of fabrics. Most penetrative is D_3, which produced northeast–plunging, open to close F_3 folds and axial-planar S_3 crenulation cleavage. The S_3 orientation is subparallel to the D_2 instantaneous stretching direction, the expected orientation of S-fabric development, making mesoscopic interpretation of D_2 fabrics difficult due to the possible confusion between S_3 and the D_2 mylonitic S-fabric. Development of S_3 is broadly synchronous with peak metamorphism and is manifested by the growth of amphiboles parallel to S_3. This represents a possible microstructural avenue to differentiate between S_3 and the mylonitic S-fabric. Weak, conjugate, S_4 spaced foliations and open folds overprint the older fabric elements about a north-south axis.

The final deformation (D_5) produced a variety of narrow pseudotachylite veins and breccia zones. The pseudotachylite veins are narrow (1–2 mm), extremely fine grained, and have distinct chilled margins locally forming the matrix to angular breccia zones. Pseudotachylite development is subparallel to S_2 and appears restricted to the JSZ (Beaumont-Smith and Rogge, 1999; Beaumont-Smith, GS-12, this volume).

Fabric development within the gabbro is significantly different and appears less intense. The gabbro generally contains a penetrative S_2 foliation, and S_3 is locally observed. The development of discrete zones (1–3 m) of D_2 protomylonitic to mylonitic fabrics, opposed to the pervasive mylonitization of the dacite, suggests that the D_2 deformation may have been partitioned into the dacite as a result of its high silica content and the resultant diffusion-dependent strain-softening mechanisms provided by quartz-rich rocks. Discrete protomylonitic and mylonitic zones are characterized by anastomosing, fine plagioclase ribbons in the gabbroic matrix and plagioclase pressure fringes on pyroxene (now hornblende) phenocrysts, producing an augen texture (Fig. GS-13-4). Mylonitic zones are defined by dextral (σ-type) asymmetrical pressure-fringe development. The finely laminated mafic tectonite units are interpreted as recrystallized zones of intense crenulation-cleavage development (cf. Beaumont-Smith and Rogge, 1999).

The distribution of dacite and gabbro delineate tight to isoclinal F_2 folds. This accounts for the multiple dacite horizons and the great variation in the thickness of the dacite. The dacite is considerably thinned on the F_2 fold limbs, whereas the thick areas of dacite (up to 25 m) occupy F_2 hinge regions. The F_2 folds are Z-asymmetrical and plunge moderately east. This geometry has the potential to cause considerable fold thickening of the dacite, improving the economic potential. The distribution of the dacite can be explained without invoking F_1/F_2 fold interference or noncylindrical F_2 folding, and does not reflect significant F_3 refolding.

Figure GS-13-4: Discrete shear zone (fine grained, lower portion of photo) developed in coarse-grained porphyritic gabbro (upper portion of photo).

MINERALIZATION AND ALTERATION

Gold mineralization was identified in 1999 through the geochemical analysis of several widely spaced samples of sheared and weakly sericitized dacite that contained trace, finely disseminated arsenopyrite. The exact nature and distribution of the gold mineralization is poorly understood at present, but detailed sampling was carried out this season to address this situation.

Based on field observations and the limited geochemical analyses, gold mineralization appears to be associated with syn-shear silicification and sericitization of the host dacite, and the presence of finely disseminated arsenopyrite. The silicification resulted from the emplacement of quartz ribbons and veins during D_2 mylonitization, and the degree of silicification ranges from simple quartz-vein emplacement (Fig. GS-13-5) to intense silicification producing massive, cherty regions in the dacite. The sericitization appears to have accompanied the silicification, with S_2 defined by muscovite-rich foliation septa. The muscovite is beige to pale green in colour and locally imparts a greenish hue to the dacite. The green tint may represent a fuchsitic component to the muscovite, with the required chromium having been liberated from the adjacent gabbro.

Extremely finely disseminated arsenopyrite and lesser pyrite, associated with the gold mineralization, form zones of dark reddish brown bloom on the weathered surface of the dacite. The distribution of the bloom corresponds to zones of silicification in sericitized dacite, either as quartz veins or bulk silicification, that are locally several metres in width and continuous along strike for tens of metres. The association between gold mineralization and presence of the sulphide minerals is reinforced by the absence of arsenopyrite in unaltered dacite and the adjacent gabbro.

Further investigations into the nature of this gold mineralization include detailed geochemical analysis of the host dacite and intercalated gabbro to determine the distribution of gold in these rocks, and the characterization of the associated alteration. Microstructural analysis will determine the relationship between the mineralizing event and the deformation, and micro-analytical analysis of mineralized samples will investigate the distribution of gold and the suspected link to sulphide concentration and species.

CONCLUSIONS

Gold mineralization in the west Gemmell Lake area is associated with the presence of finely disseminated arsenopyrite in sericitized and silicified, dacitic volcaniclastic rocks within the Johnson Shear Zone. The mineralization and accompanying alteration are the result of syn-D_2 shear-zone development, which is characterized by intense foliation and Z-asymmetrical, tight to isoclinal folding. This gold mineralization, hosted by felsic volcaniclastics, associated with ansenopyrite and coeval with shear zone development, represents a previously unrecognized style of mineralization associated with the Johnson Shear Zone.

ACKNOWLEDGMENTS

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Figure GS-13-5: Moderately sericitized and silicified volcaniclastic dacite with late-stage D_2 quartz veins (arrow) and D_2 dextral shear bands (SB).

Peck, D.C., Lin, S., Atkin, K. and Eastwood, A.M. 1998: Reconnaissance structural studies of the Au metallotects in the Lynn Lake greenstone belt (parts of NTS 64C/10, 11, *15*); *in* Report of Activities 1998, Manitoba Energy and Mines, Geological Services, p. 69–74.

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GS-14 PRELIMINARY STRUCTURAL AND GOLD METALLOGENIC STUDIES AT THE BURNT TIMBER MINE AND SURROUNDING AREA, LYNN LAKE GREENSTONE BELT (NTS 64C/10)

by L.R. Jones[1], C.J. Beaumont-Smith and B. Lafrance[1]

Jones, L.R., Beaumont-Smith, C.J and Lafrance B. 2000: Preliminary structural and gold metallogenic studies at the Burnt Timber mine and surrounding area, Lynn Lake greenstone belt (NTS 64C/10); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 69-72.

SUMMARY

The Burnt Timber deposit is one of several gold deposits along the Johnson Shear Zone, a major east-trending fault zone on the southern margin of the Lynn Lake greenstone belt in northern Manitoba. The deposit is hosted by mafic tectonite of the Wasekwan Group. Gold is associated with silica, pyrite and carbonate enrichment that occurred during dextral shearing along the shear zone.

The Johnson Shear Zone has a strong east-trending shear foliation. This foliation is deformed by chevron folds, which are, in turn, sheared along the shear foliation, forming isolated fold packets in the shear zone. These structures are overprinted by a crenulation cleavage and kinks that formed during later dextral shearing. East-trending cataclasite units and pseudotachylite veins cut across the earlier ductile structures, and are displaced by late, north-trending, sinistral faults.

INTRODUCTION

The Burnt Timber deposit is located 13 km southeast of the town of Lynn Lake (Fig. GS-14-1). Open-pit mining of the deposit began in 1993, and the mine produced approximately 80 000 ounces of gold during the following three years (Richardson and Ostry, 1996). The deposit is hosted by strongly sheared mafic tectonite, which underwent strong silicification, carbonatization and pyritic alteration over a width of approximately 100 m within the 250 m wide Johnson Shear Zone (JSZ). The JSZ is a regional dextral shear zone that predates peak regional metamorphism but cuts across regional F_1 folds of the Lynn Lake belt (Beaumont-Smith and Rogge, 1999). A late brittle fault, the T1 fault, is subparallel to the JSZ. The T1 fault is a north-dipping reverse fault (Peck et al., 1998). The deposit is in the hanging wall of the fault and extends approximately 40 m from the fault.

Because the shear zone has a complex deformation history, involving both dextral shearing and brittle faulting, the timing of gold mineralization and the geometry of the gold-bearing structures have not yet been resolved. Following a preliminary study by Beaumont-Smith and Rogge (1999), the present study was initiated to document the relationships between the structures, alteration and gold mineralization of the Burnt Timber deposit and surrounding area.

ROCK UNITS

The Burnt Timber deposit is hosted by mafic volcanic rocks of the Wasekwan Group (Bateman, 1945). The rocks are fine-grained, dark green, mafic tectonite, in which all primary structures have been erased by the strong deformation in the shear zone. Near the deposit, the mafic tectonite has alternating dark grey and light grey bands that weather pink and green, respectively, on outcrop surface. The mafic tectonite is locally strongly magnetic due to the presence of magnetite crystals up to 2 mm in size.

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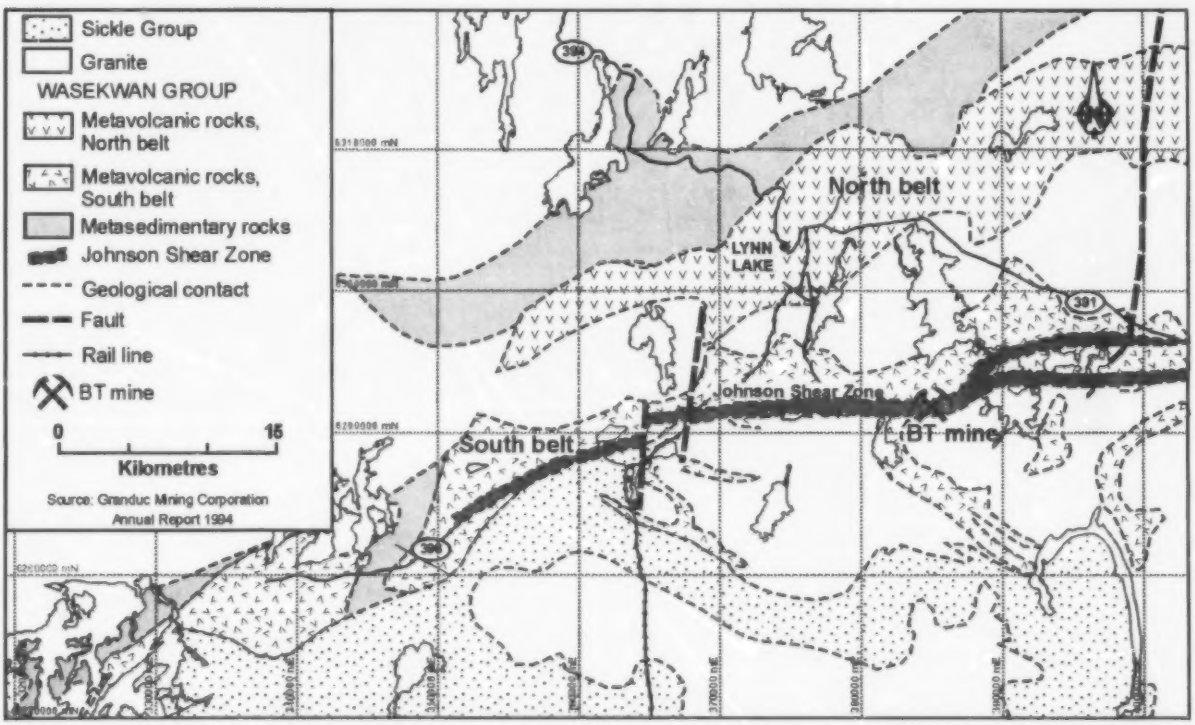

Figure GS-14-1: Simplified geological map of the Lynn Lake belt, showing the location of the Johnson Shear Zone and study area.

To the north and south of the JSZ, the mafic volcanic rocks are fine-grained massive flows with few preserved quartz-filled amygdules. The mafic flows are dark green on fresh surface and light green on outcrop surface. Minor units of mafic volcaniclastic rock and felsic tuff are interlayered with the mafic flows. The mafic volcaniclastic rocks are exposed for more than 100 m across strike. They are finely laminated with alternating, light grey and dark grey layers less than 0.5 cm wide. Younging indicators are rare, but graded bedding on two outcrops suggests that the beds young to the north. To the north of the JSZ, felsic tuff occurs as thin layered units, which are typically less then 4 m thick. The tuff is white on outcrop surface, light grey on fresh surface, and finely laminated to thickly bedded (10 cm maximum thickness). The felsic tuff is cut by a cleavage that refracts toward the bedding as it passes from the more massive thicker beds to the thinner, presumably less competent beds.

South of the deposit, the mafic metavolcanic rocks are cut by a coarse-grained granodiorite of the Pool Lake suite (Gilbert et al., 1980). The intrusion comprises 25% quartz, 60% feldspar and 15% biotite. Biotite is altered to chlorite in the margins of late fractures cutting through the intrusion. North of the Burnt Timber deposit, a poorly exposed, dark grey gabbro contains large amphibole grains up to 2 cm in size. The granodiorite and gabbro are not in contact, so the relative age of the two intrusions is not known.

STRUCTURE

Three of the five deformational events that affected the metavolcanic rocks in the study area are associated with movements along the JSZ. The JSZ formed during the second deformational event (D_2). Dextral shearing during D_2 produced a strong shear foliation (S_2). Dextral shear-sense indicators include 1) asymmetrical, back-rotated quartz boudins; 2) intrafolial Z-folds (Fig. GS-14-2); 3) dextral dragging of S_2 along the margin of competent rock units; 4) dextral offsets of markers; and 5) asymmetrical strain shadow around competent silicified volcanic fragments (Fig. GS-14-3). Near the middle of the JSZ, S_2 is folded by cascades of dextral F_2 chevron folds (Fig. GS-14-4), which are locally sheared by S_2, forming isolated fold packets surrounded by S_2. Ridge-and-groove lineations along S_2 are steeply plunging (60–70°), suggesting that the shear zone underwent dip-slip as well as strike-slip movements.

An early D_1 structure is only observed in the granodiorite intrusion. The intrusion is cut by a penetrative S_1 foliation defined by biotite and quartz ribbons. The S_1 foliation is in turn overprinted by narrow shear zones parallel to the JSZ. This foliation is likely preserved within the more competent granodiorite body because of the partitioning of the D_2 deformation in the less competent metavolcanic rocks.

The third deformational event (D_3) produced dextral F_3 kinks, which overlap in style with the F_2 chevron folds (Beaumont-Smith and

Figure GS-14-2: D_2 chevron folds folding the S_2 foliation in the Johnson Shear Zone.

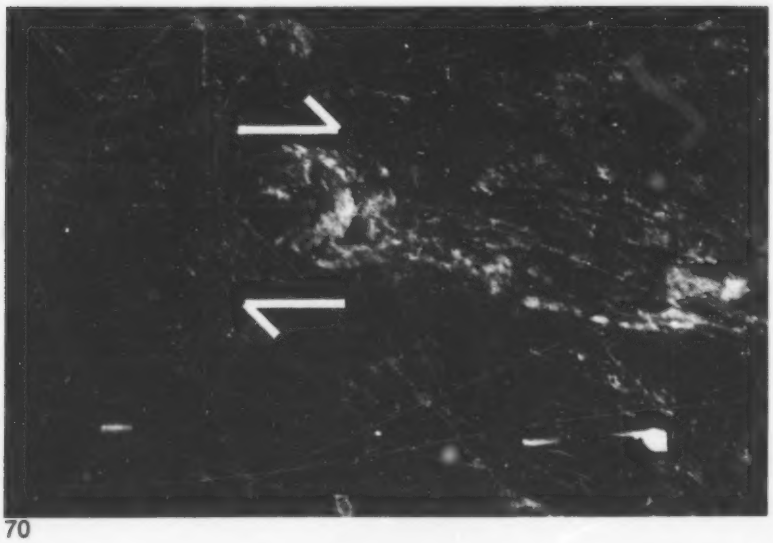

Figure GS-14-3: Asymmetrical Z-fold formed by D_2 dextral movement along the Johnson Shear Zone.

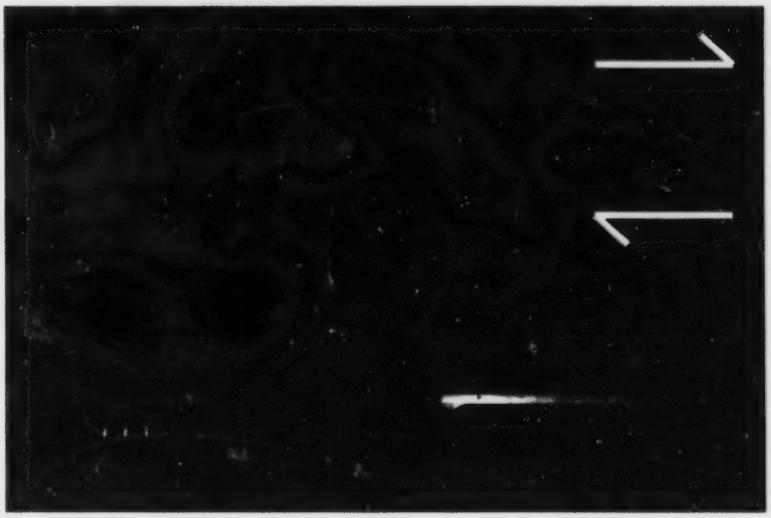

Figure GS-14-4: Dextral asymmetrical strain shadows around competent volcanic fragments in the Johnson Shear Zone.

Rogge, 1999). A crenulation cleavage (S_3) is axial planar to the F_3 folds. The D_3 kinks and folds strike northeast, slightly counterclockwise of F_2 axial planes. The orientation of the F_3 kinks and their overlap in style with F_2 folds suggest that both structures formed during progressive (D_2-D_3) dextral shearing along the JSZ.

The JSZ is overprinted by brittle structures related to two distinct deformation events (D_4 and D_5). The D_4 faults strike east-west and dip steeply to the north. Narrow cataclastic zones, consisting of very small (<1 mm) angular rock fragments and dark pseudotachylite veins with quenched margins, are formed during D_4 faulting. The offset of quartz veins and the dragging of S_2 along minor faults suggest that the faults are dextral. The D_5 faults trend north, dip to the west and clearly displace D_4 pseudotachylite units (Fig. GS-14-5). Where both D_5 and D_4 structures are present on outcrop, the mafic tectonite has been strongly disrupted, as shown by abrupt and discontinuous changes in the trend of S_2 and quartz veins.

The T1 fault cuts across the east- and west-facing walls of the Burnt Timber open pit. The open pit is now flooded, so the fault could not be examined. The fault orientation and reverse slip are consistent with both D_4 and D_5 faulting events, so the fault could have formed during either event.

GOLD MINERALIZATION

Quartz-carbonate alteration of the mafic tectonite occurs across the JSZ. Quartz and iron carbonate veins are folded by F_2 chevron folds and overprinted by the sericitic S_2 shear foliation. In high-grade zones of the orebody, the mafic tectonite underwent strong silicification, pyritization and carbonatization. Sericite along S_2 foliation planes is replaced by chlorite. This supports the conclusions of a geochemical study by Peck and Eastwood (1996), who reported an inverse correlation between gold and potassium values in the orebody.

A spatial relationship between felsic dykes and gold mineralization is observed in several drillholes through the deposit. Gold assay values increase where felsic dykes are present. This suggests a possible genetic association between the dykes and the gold mineralization.

CONCLUSIONS

Mafic metavolcanic rocks in the vicinity of the Burnt Timber mine underwent five deformational events. The JSZ is a regional, dextral, D_2 shear zone. It is overprinted by D_3 kinks and folds, which crenulate and kink the S_2 shear foliation. The final two brittle events (D_4 and D_5) produced cataclasite and pseudotachylite veins, which offset the older ductile structures.

Gold mineralization is associated with quartz-carbonate-pyrite

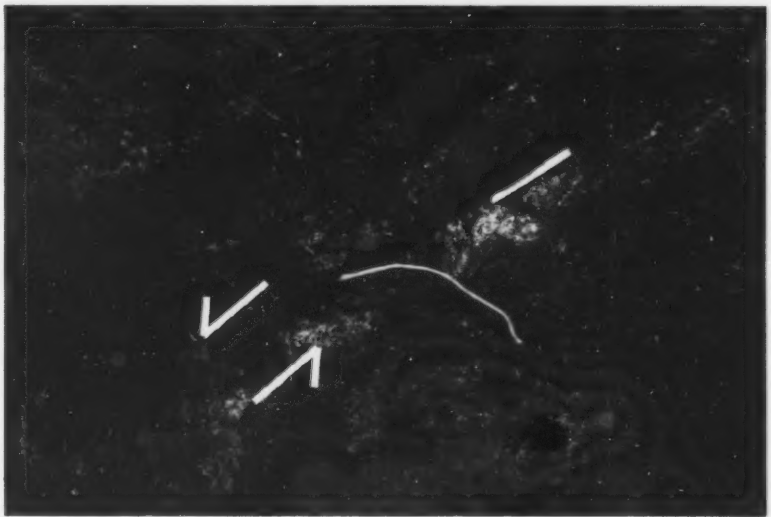

Figure GS-14-5: D_4 pseudotachylite offset by D_5 brittle faults.

alteration of the mafic tectonite in the JSZ. There is also a spatial relationship between gold mineralization and felsic dykes, which may become a key prospecting tool for finding new deposits along the JSZ.

Future work will involve whole-rock geochemistry of the altered mafic tectonite hosting the Burnt Timber deposit, and detailed structural mapping of other gold occurrences along the JSZ.

ACKNOWLEDGMENTS

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GS-15

STRUCTURAL ANALYSIS AND GOLD METALLOGENY OF THE FARLEY LAKE GOLD DEPOSIT, LYNN LAKE GREENSTONE BELT (NTS 64C/16)

by C.J. Beaumont-Smith, D.R. Lentz[1] and E.A. Tweed[2]

Beaumont-Smith, C.J., Lentz, D.R. and Tweed, E.A. 2000: Structural analysis and gold metallogeny of the Farley Lake gold deposit, Lynn Lake greenstone belt (NTS 64C/16); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 73-81.

SUMMARY

Preliminary structural and geochemical investigations of the Farley Lake gold deposit have determined that the gold mineralization is the result of emplacement of postdeformational, postplutonic quartz-carbonate-sulphide veins into banded oxide- and silicate-facies iron-formation, and the accompanying sulphidization of the iron-formation. The high-grade quartz-carbonate-sulphide veins were emplaced along pre-existing, shallowly southwest-dipping joints and subvertical D_2 faults and shears. The thick sequence of iron-formation and interbedded argillaceous sedimentary rocks that hosts the mineralization has undergone significant D_1 and D_2 fold thickening.

INTRODUCTION

The Farley Lake gold deposit is located in the northern Lynn Lake greenstone belt, approximately 45 km northeast of the town of Lynn Lake (Fig. GS-15-1). The deposit consists of four zones of gold mineralization, the Wendy, South, Southeast and East zones (Richardson and Ostry, 1996), hosted by a thick sequence of banded oxide-, silicate- and sulphide-facies iron-formation (Fedikow, 1995). Production from the Wendy and East open pits between 1996 and 1999 totalled, based on published reserves, 635 000 t of pyrrhotite-associated ore averaging 6.86 g/t Au and 363 000 t of pyrite-associated ore averaging 4.80 g/t Au, respectively (Richardson and Ostry, 1996). The total preproduction resource stood at 1.45 million t grading 3.90 g/t Au (Richardson and Ostry, 1996).

The Farley Lake deposit is hosted by Wasekwan Group (Bateman, 1945) volcanic and sedimentary rocks that constitute the Agassiz Metallotect (Fedikow and Gale, 1982), a stratigraphic succession defined, from top to bottom, by high Mg-Cr basalt (picrite), rhyolitic to dacitic volcanic rocks, argillaceous sedimentary rocks and exhalative sedimentary rocks. The mine sequence is tightly folded (F_1) into an overall upright anticlinal structure. A smaller synclinal inlier of carbonaceous shale, plunging moderately to the west (Peck et al., 1998), occurs in the roof of this larger F_1 feature; the anticline is therefore cored by banded iron-formation. The deposit is unique relative to other gold deposits in the Lynn Lake greenstone belt in that gold mineralization (centimetre- to decimetre-scale quartz-sulphide veins (QSV) hosted in banded iron-formation) is associated with late faults and joints that are discordant with the earlier tectonic fabrics. Other gold deposits, namely the Burnt Timber and MacLellan deposits, are characterized by the emplacement of gold mineralization and associated hydrothermal alteration early in the deformational history, possibly prior to and coincident with fabric development and peak amphibolite-grade regional metamorphism.

Investigations of the gold mineralization at the Farley Lake deposit involved detailed structural analysis of the East pit, including geochemical sampling of mineralization exposed in the pit and selected diamond-drill core. Mining operations in the Wendy pit were suspended in 1998 and the pit is completely flooded, precluding it from structural analysis. Production from the East pit concluded in October of 1999 and the pit is in the process of flooding. Structural analysis of the East pit was initiated in September 1999, with the mapping of the lower production benches. Mapping and sampling of the upper benches and preliminary diamond-drill core study and sampling were completed this summer.

DEPOSIT GEOLOGY

The Farley Lake gold deposit is hosted by a fold-thickened sedimentary succession of thinly layered iron-formation and argillaceous sedimentary rocks located within the core of an east-trending anticline (Fig. GS-15-2; Peck et al., 1998). This sedimentary sequence is overlain

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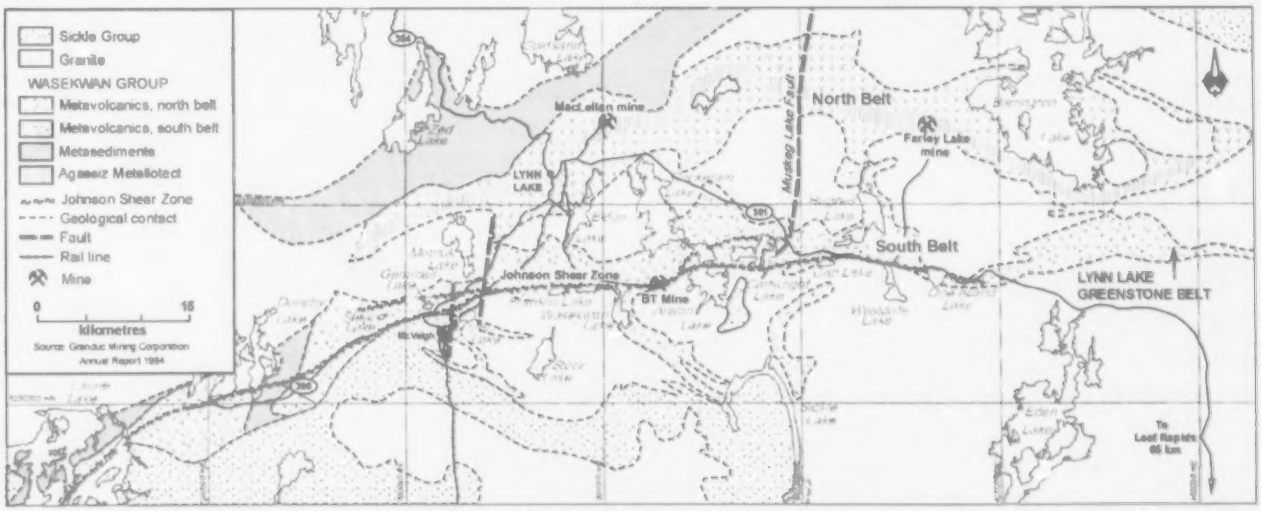


Figure GS-15-1: General geology of the Lynn Lake greenstone belt and location of the Johnson Shear Zone (after Peck et al., 1998).

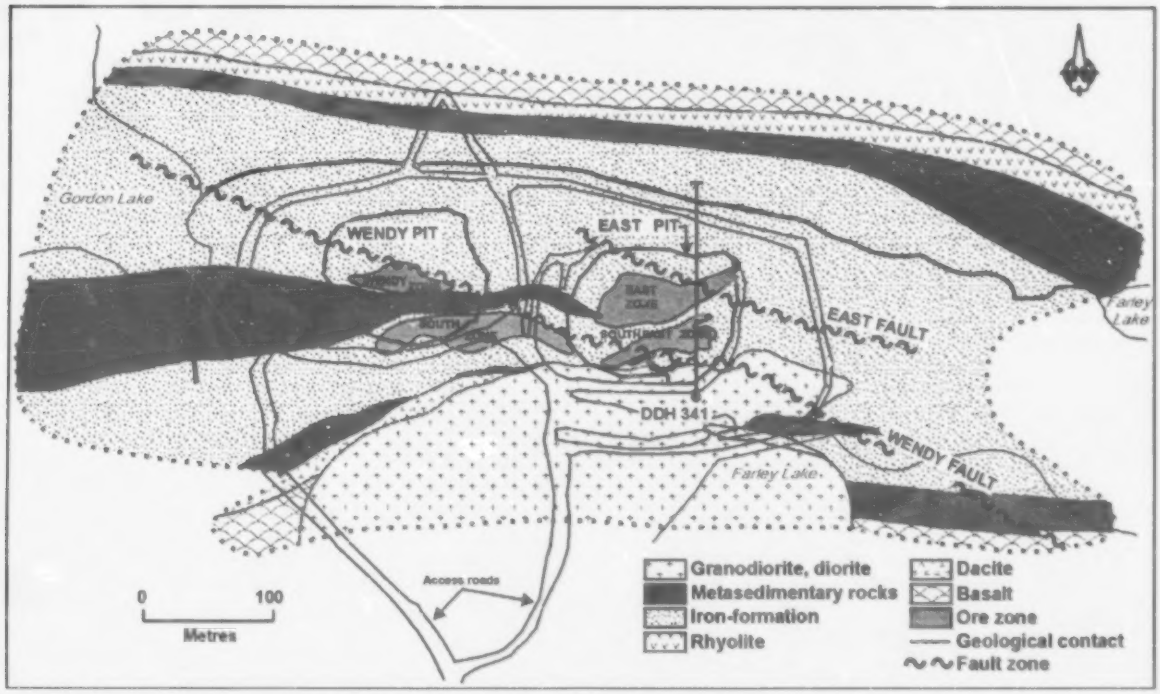


Figure GS-15-2: Local geology of the Farley Lake deposit (after Peck et al., 1998).

by dacitic (south) and rhyolitic (north) volcanic rocks and high-Mg-Cr basalt (picrite). The southern margin of the deposit has been discordantly intruded by a granodioritic to dioritic pluton of the Pool Lake intrusive suite (Gilbert et al., 1980). The pluton is in direct contact with mineralized iron-formation along the southern edge of the East pit.

The iron-formation unit is approximately 600 m wide and has a strike length in excess of 6 km (Milligan, 1960). It is composed of, in decreasing order of abundance, finely laminated magnetite-chert (oxide facies), magnetite-grunerite-chert (silicate facies) and pyrrhotite-chert (sulphide facies). The oxide facies consists of thin, alternating magnetite and chert laminations of roughly equal proportions. The minor amounts of interlayered carbonate within this facies appears to represent transposed pre- to syn-D_2 veins. The silicate facies is composed of finely laminated chert and beige amphibole (grunerite-anthophyllite)-magnetite layers. This facies probably resulted from an increased hemipelagic component during deposition. The sulphide facies constituted a relatively minor component in the East zone of the deposit, but represents a volumetrically important constituent of the Wendy zone, largely hosting the mineralization (Richardson and Ostry, 1996). Sulphide-facies iron-formation exposed in the lower portions of the East pit comprise thin (1–2 m) interbedded units of finely laminated chert with thin pyrrhotite laminae and abundant, coarsely disseminated pyrrhotite porphyroblasts. The latter form 1 to 2 mm blebs and trains of discontinuously interconnected blebs. Where well foliated, thin pyrrhotite stringers define the foliation. Total pyrrhotite content approaches 40% by volume.

There appears to be a systematic distribution of oxide and silicate iron-formation facies in the East pit. The southern portion of the pit is dominated by oxide facies with minor interbeds of silicate- and sulphide-facies iron-formation. The northern portion of the pit is underlain by silicate-facies iron-formation with only minor oxide facies.

Interbedded with the iron-formation are thin, clastic sedimentary units (approx. 1–2 m average thickness), which locally reach a thickness of 50 m in the core of the regional antiform. These rocks comprise thinly bedded sequences of argillite, chloritic argillite and minor siltstone. Several 10 to 20 cm thick silicate-facies grunerite-garnet iron-formation horizons are tightly fold-intercalated with the thick argillite unit exposed in the west wall of the East pit.

Recrystallization associated with regional lower amphibolite facies metamorphism has resulted in the general coarsening of the constituent mineral phases, the growth of amphibole in the silicate-facies iron-formation, and the development of calc-silicates in local calcareous iron-formation and premetamorphic carbonate veins. The timing of metamorphism is broadly coeval with the third phase of regional deformation (D_3) and postdates the intrusion of the Pool Lake suite. The growth of randomly oriented grunerite porphyroblasts overprints the main S_2 foliation. Locally, the porphyroblasts have a strong S_3-parallel preferred orientation, suggesting that the differences in degree of porphyroblast preferred orientation reflect slight differences in the timing of grunerite growth due to slight changes of bulk rock chemistry.

Intruding the Wasekwan supracrustal succession is a Pool Lake suite granodioritic to dioritic pluton. The pluton is medium grained, generally equigranular and locally well foliated, particularly near the northern contact with the supracrustal sequence. Although commonly referred to as granodioritic, the intrusion seems to have several phases that were recognized during the original drilling program; the outermost phases are tonalitic to dioritic in composition (*see* next section). Trace to several per cent pyrrhotite and pyrite are commonly found with quartz and carbonate veining associated with coincident chloritized, silicified and carbonatized zones. Although locally mineralized with gold-bearing sulphide-quartz (± carbonate±chlorite) veins, there are numerous late fabrics with sulphide-quartz veins and alteration within each of the intrusive phases that are devoid of gold mineralization. The alteration zones, which are locally associated with anomalous gold abundances, are intimately associated with ductile to brittle shearing that seems to overprint a pre-existing ductile foliation defined by the preferred orientation of biotite, amphibole and plagioclase (i.e. peak metamorphic assemblages). In general, the fabric intensity (weak to intense) and alteration increase toward the northern margin of the pluton. Chloritization of the constituent biotite and hornblende, together with sericitization of the feldspar, results in a distinct colour change that seems to coincide with the introduction of quartz and carbonate, as well as pyrite and pyrrhotite. It is uncertain whether there was considerable early-phase

syndeformational metasomatism, especially since the pluton seems to be reversely zoned toward the margin. The relative timing of the granodioritic and tonalitic phase to the dioritic phase is uncertain. The S_2 foliation seems most intense in the outermost zones of the intrusion; there is minimal fabric within the granodioritic phase, even though there is a high percentage of ferromagnesian silicates present (25%). Nonetheless, the first-phase intrusions predate the main (F_2) folding event, although they are discordant with the folded stratigraphy. This indicates that subsequent deformation, coincident with peak amphibolite-grade metamorphism, followed their intrusion.

GRANODIORITIC TO DIORITIC INTRUSION

In order to ascertain whether the compositional change from the granodiorite to the intensely foliated intrusive margin is associated with a specific episode of hydrothermal alteration or is related to a primary compositional effect, six drill-core samples were taken from available core (diamond-drill hole DDH 341) that was drilled from south to north into the southeastern Farley Lake zone (East pit area). The six-sample profile has an increase in ferromagnesian phases (biotite±amphibole), in part retrograded to hydrothermal biotite, then chlorite and carbonate, from the weakly altered granodiorite into the diorite along the northern contact. Petrographically, the intrusive phases obviously exhibit more intense hydrothermal alteration of the lower temperature variety (Mg-chlorite–carbonate–sulphide) approaching the northern contact. However, granodioritic phases exhibit considerable interstitial biotitic alteration, which seems to have replaced an earlier hornblende- and biotite-bearing granodioritic assemblage. The secondary biotite seems to have a deeper reddish hue than the green igneous biotite. Titanite seems to have been formed as a result of the selective ferromagnesian replacement processes, although coupled with plagioclase alteration (anorthite replacement). The gold-enriched sample (595 ppb Au at 152.4 m [500 ft.] in DDH 341) has intergrown Mg-rich chlorite and carbonate replacing reddish biotite. Pyrrhotite with very minor chalcopyrite occurs in the groundmass with the secondary assemblages, although it is unclear if it is associated with chlorite-carbonate or the hydrothermal biotite. The presence of chalcopyrite in the gold-bearing section is surprising, considering the sample only contained 85 ppm Cu. Based on the original drill log, the altered diorite has an apparent width of approximately 18.3 m (60 ft.) or a 5.2 m (17 ft.) true width. The most intensely altered sample (103.9 m [675 ft.] in DDH 341) contained minor remnant clinopyroxene replaced by an Fe-Mg amphibole (fine to coarse grained), which was in turn partially replaced by and intergrown with Mg-chlorite. The amphibole has a weak preferred orientation. Rutile is secondary after ilmenite. There are several late carbonate veinlets in this sample.

These sample cores were crushed and pulverized in a soft-iron swing mill by the Manitoba Geological Survey and analyzed for major and trace elements, by x-ray fluorescence spectroscopy, instrumental neutron activation analysis and inductively coupled plasma–mass and optical emission spectroscopy, at Activation Laboratories Ltd., Ancaster, Ontario (Table GS-15-1). For duplicate elements, preference was given to techniques not involving dissolution of samples.

Figures GS--15-3 and -4 are variation diagrams illustrating the major-element composition of samples across the profile. As can be seen in Table GS-15-1, there is a consistent decrease in SiO_2 from the core toward the outer margin of the intrusion, consistent with petrographic observations. The Al_2O_3 content decreases slightly from 15.76 to 15.35 wt % with decreasing SiO_2, but then drops off to 8 wt % in the intensely altered (amphibolitized and chloritized) zone near the intrusion margin. If this is a diorite, it is possible to have lower Al_2O_3 contents with higher $Fe_2O_{3\ tot}$ and MgO (32.1 and 9.3 wt %, respectively) at the margin (Fig. GS-15-3b, -3c). The K_2O, Na_2O and CaO contents decrease markedly toward the margin as well (Fig. GS-15-3d, -3e, -3f), which is a function of both primary composition and alteration. The increase in ferromagnesian components at the expense of alkali and alkali-earth elements, including Rb, Ba and Sr, and decreased SiO_2 content are consistent with increasing amphibolitization (?cummingtonite) and chloritization toward the outer margin. The original drill logs noted that these rocks were granodioritic, tonalitic and dioritic in mineralogy, based on textures, many of which were relict. Therefore it is difficult to ascertain whether the more mafic looking rocks, which have greater fabric development and more intense alteration, are actually tonalitic and dioritic in composition. It is possible to test this using immobile elements and their ratios.

Figure GS-15-4a illustrates consistently increasing TiO_2 with decreasing SiO_2. This cannot be related to mass changes, as it is opposite to what would be expected based on the decrease in Al_2O_3 (8 wt %). Also, the TiO_2 co-varies strongly with Sc and V, consistent with an iron tholeiitic composition for the diorite, whereas the other granodioritic and tonalitic samples have a more obvious calc-alkalic signature. There is a decrease in P_2O_5, S, Ni (Fig. GS-15-4b, -4c, -4d), Co and Cr at the outer margin relative to the less altered tonalitic and granodioritic phases of the intrusion. The primitive-mantle–normalized spider diagram (Fig. GS-15-5) for these samples shows that the profiles for the granodiorite and tonalite are similar, although the dioritic sample is notably lower in incompatible elements, except the heavy rare-earth elements (HREE), and higher in TiO_2 (Sc, V), as mentioned earlier. Therefore the light rare-earth element (LREE)/HREE profile of the diorite is notably flatter, typical of a tholeiitic parentage, than for the calc-alkaline samples. Similarly, immobile-element ratios (Fig. GS-15-6a) illustrate that the dioritic sample has a typical tholeiitic signature ($Zr/Y = 3$), whereas the other samples have values between tholeiitic and calc-alkalic compositions. The Zr/TiO_2 ratio is also consistent with a gabbroic to dioritic composition. All samples plot in the I-type granitoid (volcanic-arc–related) field using the Nb-Y discrimination diagram (Fig. GS-15-6b; Pearce et al., 1984), consistent with their generally aluminous composition. The Th-Zr-Nb discrimination diagram (Fig. GS-15-6c; Wood et al., 1979) shows that the granodioritic and tonalitic samples are of volcanic-arc affinity, whereas the dioritic sample has an enriched mid-ocean ridge basalt (MORB) to within-plate association, which is consistent with the Fe, Mg and TiO_2 contents. The Zr versus Ga/Al contents of these samples (Fig. GS-15-6d; Whalen et al., 1987) indicate an arc signature for the granodiorite and tonalite, although the diorite falls into the anorogenic field, which fits with the tholeiitic (proto-arc or rift-related) interpretation, but not with the Nb/Y ratio, which is much less than 0.7 (*see* Winchester and Floyd, 1977). Based on textural and compositional considerations, it is probable that the diorite is an earlier intrusive phase that was followed by the main volcanic-arc–associated tonalite and granodiorite.

It is also notable that the sample enriched in gold (596 ppb) is associated with high sulphur, although it is evident from the logs and gold assays that this is not always the case. Also, the one anomalous gold sample has low As, Sb, Mo and W. Interestingly, Fedikow (1995) indicated that the gold was not associated with some of the typical indicator elements at the Wendy zone. The sample with the highest gold content is not associated with the most altered (chloritic) sample.

STRUCTURAL ANALYSIS

A major objective of this study is to determine the timing of mineralization relative to alteration, deformation, metamorphism and plutonism in the area. Therefore, an understanding of the structural history of the deposit is critical. Detailed structural analysis of the East pit has delineated four generations of folds (F_1–F_4). The geometry and distribution of the iron-formation appear to be largely controlled by F_1 and F_2 folds, with overprinting by the two younger folding events producing minor redistribution of the older fabric elements. Mesoscopic F_1 folds are rare and are generally recognized as isoclinal, commonly intrafolial folds overprinted by S_2 or refolded by upright, tight, F_2 chevron folds (Fig. GS-15-7). The F_2 folds plunge steeply to the east and are dominantly Z-asymmetrical. A ubiquitous S_2 axial-planar foliation

Table GS-15-1: Analytical data for drillcore samples from the Pool Lake suite granodioritic to dioritic pluton, southeastern Farley Lake zone.

	Sample number (and rock type)					
	DDH341-100 (granodiorite)	**DDH341-300 (granodiorite)**	**DDH341-500 (tonalite)**	**DDH341-515 (tonalite)**	**DDH341-580 (tonalite)**	**DDH341-675 (diorite)**
Major elements (wt %):						
SiO_2	56.51	55.87	53.92	50.34	45.91	40.16
Al_2O_3	15.76	15.88	15.46	15.14	15.35	8
Fe_2O_3	5.93	6.69	6.97	7.8	10.61	32.1
MnO	0.102	0.093	0.117	0.104	0.18	0.361
MgO	3.39	4.05	4.03	5.56	8.8	9.3
CaO	5.86	5.99	6.58	6.39	4.81	2.83
Na_2O	4.93	4.48	3.83	4.31	3.79	0.13
K_2O	1.51	1.89	1.65	2.09	0.31	0.49
TiO_2	0.525	0.588	0.598	0.766	0.819	1.786
P_2O_5	0.34	0.38	0.42	0.5	0.26	0.09
LOI	3.73	2.85	5.15	6.06	8.69	3.9
Total	**98.59**	**98.76**	**98.72**	**99.07**	**99.52**	**99.14**
Trace elements (ppm):						
Nb	11	11	9	11	4	6
Zr	127	128	87	101	65	49
Y	11	12	15	15	22	18
Sr	983	1001	573	781	285	28
Rb	25	34	29	50	3	14
Ba	608	695	697	693	67	93
Ga	17	17	18	17	13	22
S	120	120	1790	95	300	50
Trace elements (ppm):						
Au	2	2	595	27	2	2
As	2.6	3	5.8	11.5	11	43.3
Ba	550	580	660	590	200	210
Co	15	20	23	27	48	71
Cr	80	94	107	143	47	127
Cs	11	12	9	21	2	6
Ni	49	58	57	88	110	28
Sb	0.5	0.4	0.3	0.4	0.1	0.5
Sc	10.2	12	14.2	16.3	28.3	22.4
Ta	0.5	0.5	1.4	0.5	0.5	0.5
Th	3.7	3.1	2.5	2.3	1.1	0.7
U	2.3	1	2.2	1.6	0.9	0.5
W	1	1	7	1	3	1
Zn	70	100	84	100	104	81
La	35	32.6	26.3	27.6	17.1	9.9
Ce	62	60	53	55	34	23
Nd	30	31	31	32	19	16
Sm	6	6.4	6.2	6.7	3.8	4.4
Eu	1.8	1.8	1.8	1.8	1.1	0.7
Tb	0.5	0.8	0.5	0.5	0.5	0.6
Yb	1	1.1	0.9	0.9	2	1
Lu	0.15	0.16	0.15	0.14	0.3	0.14
Trace elements by ICP-OES (ppm):						
Ag	0.3	0.3	0.3	0.3	0.3	0.3
Cd	0.3	0.3	0.3	0.3	0.3	1.6
Cu	29	64	85	49	14	1
Mn	770	772	901	769	1230	2480
Mo	1	2	2	1	2	2
Ni	47	59	57	80	103	32
Pb	15	15	21	8	15	11
Zn	65	86	66	86	124	45
V	101	121	120	150	204	658
S	0.045	0.068	0.246	0.015	0.081	0.001

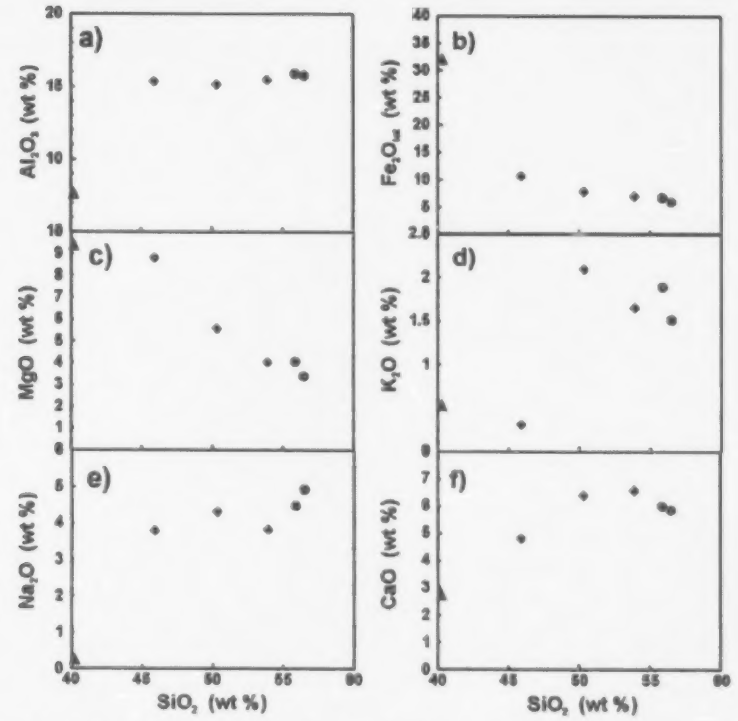


Figure GS-15-3: SiO_2 versus a) Al_2O_3, b) $Fe_2O_{3\,tot}$, c) MgO, d) K_2O, e) Na_2O, and f) CaO in the granitoid samples (see Fig. GS-15-5 for symbols).

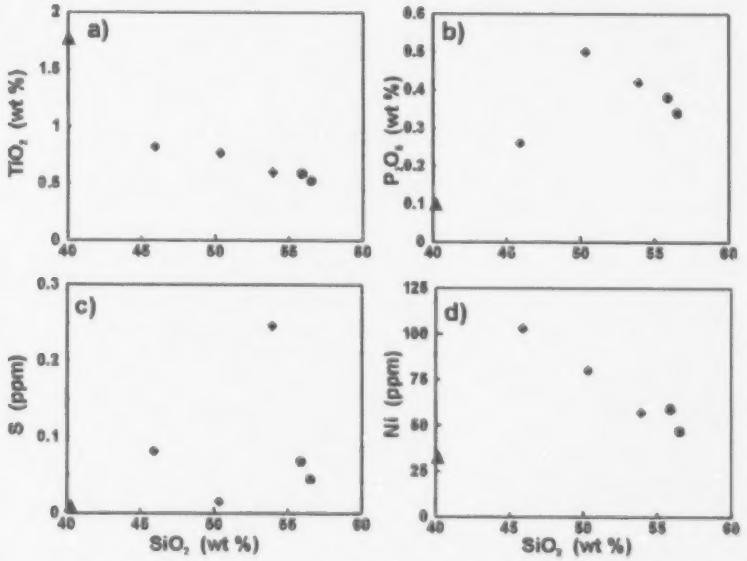


Figure GS-15-4: SiO_2 versus a) TiO_2, b) P_2O_5, c) S, and d) Ni in the granitoid samples (see Fig. GS-15-5 for symbols).

accompanied F_2 development and represents the most penetrative secondary fabric element.

Peck et al. (1998) interpreted the distribution of the iron-formation within the core of a shallow west-plunging anticlinorium (*see* Fig. GS-15-2). Key to this interpretation is the correlation of the dacite exposed along the south margin of the East pit and the rhyolite north of the pit as lateral facies equivalents. Consequently, the argillite exposed in the west wall of the East pit was interpreted to occupy the core of a shallowly west-plunging synform. Based on mesoscopic overprinting relationships, this fold is most likely a macroscopic F_1 fold. There is no systematic change in F_2 asymmetry across the argillite, suggesting that F_2 is not responsible for the distribution of the argillite and iron-formation.

The intensity of D_2 fabric development, demonstrating the partitioning of the deformation, is reflected in the large competency contrasts between the units that form the deposit stratigraphy. The iron-formation is tightly folded and moderately foliated. A steep D_2 strain gradient

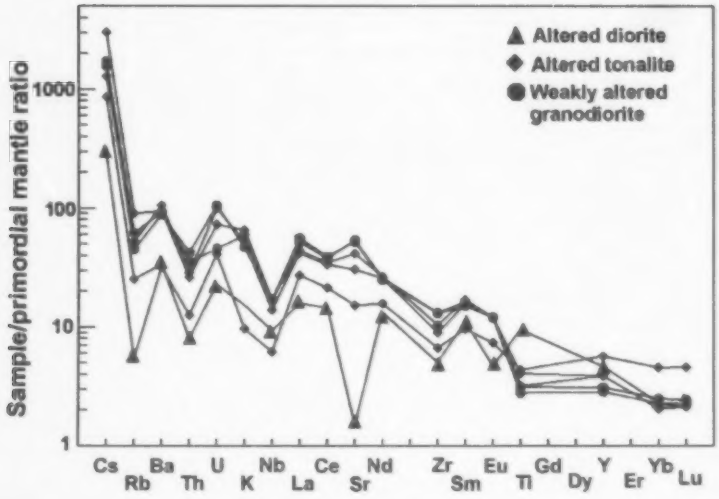


Figure GS-15-5: Primitive-mantle–normalized spider diagram with the six samples illustrated (see Sun and McDonough, 1989 for normalizing values). The bolded elements are considered immobile under mostly hydrothermal alteration conditions.

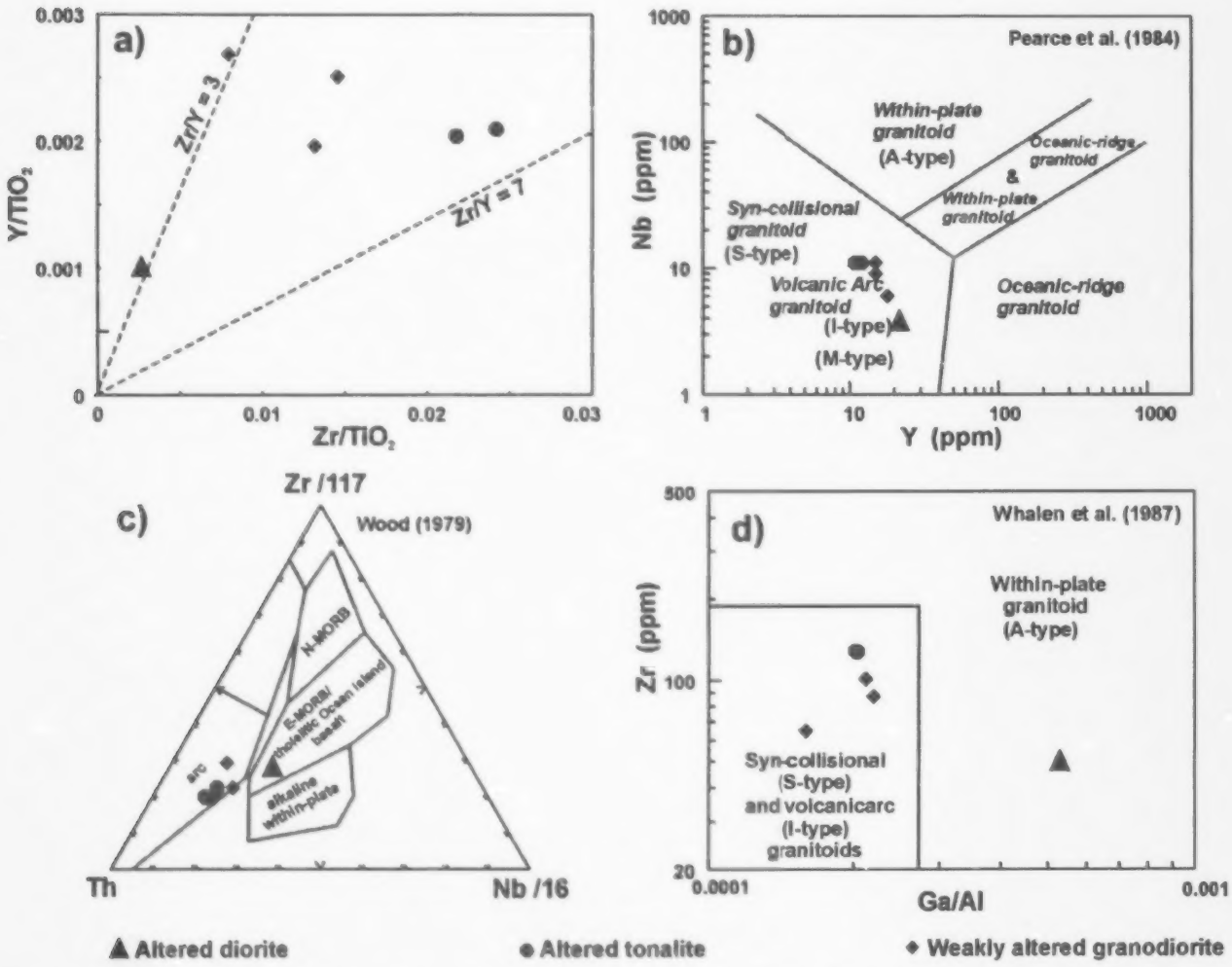


*Figure GS-15-6: a) Y/TiO_2 versus Zr/TiO_2 diagram, with Zr/Y illustrated, b) Nb versus Y discrimination diagram (*modified after *Pearce et al., 1984), c) Th-Zr-Nb discrimination diagram (*modified after *Wood, 1979); abbreviations: A, N-MORB affinity; B, E-MORB affinity; C, within-plate affinity; and D, volcanic-arc affinity, d) Zr versus Ga/Al (*see *Whalen et al., 1987).*

Figure GS-15-7: Steeply east-plunging F_2 fold in oxide-facies iron-formation.

along the southern contact between the iron-formation and the granodioritic intrusion is characterized by the development of protomylonite and a narrow zone of mylonitic fabrics in the iron-formation along the contact. The argillite interbedded with the iron-formation is intensely folded and foliated, as evidenced by the development of phyllitic zones and local fault displacement of F_2 limbs. The margin of the intermediate pluton exposed in the south wall of the East pit contains moderate S_2 foliation development and preserves an older S_1 slaty cleavage. An S_3 foliation and F_4 folds are also developed in the granodiorite.

Overprinting F_2 are two generations of close to open folds. The F_3 generation comprises northeast-trending, steeply plunging, Z-asymmetrical folds. Although mesoscopic F_3 folds are rare, a penetrative S_3 crenulation cleavage overprints older fabric elements. The youngest generation of folds comprises north- to northwest-trending, open chevron folds. Folds of the F_4 generation generally lack an axial-planar foliation, but produced small-scale tight crenulations, which locally have differentiated limbs that approximate a foliation.

Emplacement of the intermediate dykes postdates F_2 folding. The orientation of the dykes is subparallel to F_2 axial planes and the dykes do not outline any F_2 folds or boudinage. The dykes predate F_3 and generally contain an S_3-parallel spaced fracture cleavage. The intermediate intrusion (granodiorite to diorite) of the Pool Lake suite seems to intrude during late F_2 or early F_3.

Numerous premineralization faults transect the East pit. Many of these occupy F_2 limbs and appear to represent adjustments in response to the tight F_2 folding. Two major northwest-trending faults, the East and Wendy faults, postdate the folding events, since they transect F_2–F_4 axial traces. These parallel faults comprise steeply northeast dipping zones of intense fracturing cored by decimetre-scale gouge zones characterized by imbricated clasts, indicating a large component of normal movement.

GOLD METALLOGENY

Gold mineralization in the East pit is associated with two sets of centimetre- to decimetre-scale, coarse-grained, quartz-sulphide and quartz-dolomite-sulphide veins with minor chlorite (Fig. GS-15-8). The veins have a wide variety of textures, including open space, vuggy textures; laminated, crack-seal textures; and locally composite, zoned veins consisting of crack-seal textures along one side and sulphide-matrix quartz breccia along the other.

The timing of mineralization postdates both the folding and faulting events. There are numerous examples of mineralized flats transecting F_4 fold axes, the latest folding event, and both the Wendy and East faults (Fig. GS-15-9a, -9b). The shallow QSV (flats) were emplaced along shallowly southwest-dipping faults and joints. These brittle-ductile features record a sinistral offset of approximately 1 m. Narrow zones of sinistral ductile bending of the bedding laminations indicate the initiation of the joints as ductile structures followed by brittle failure due to the elevation of the strain rate. The subvertical QSV (steeps) were emplaced along moderate to steep south-southwest-dipping joints, along subvertical S_2 parallel faults, and within zones of intense S_2 foliation development (phyllitic zones). These mineralized veins are very high grade (approx. 30 g/t Au; Peck et al., 1998) and are mantled by zones (up

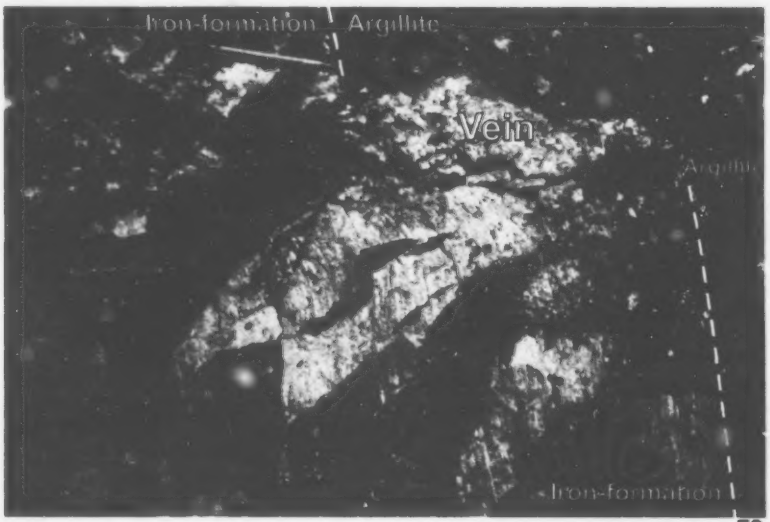
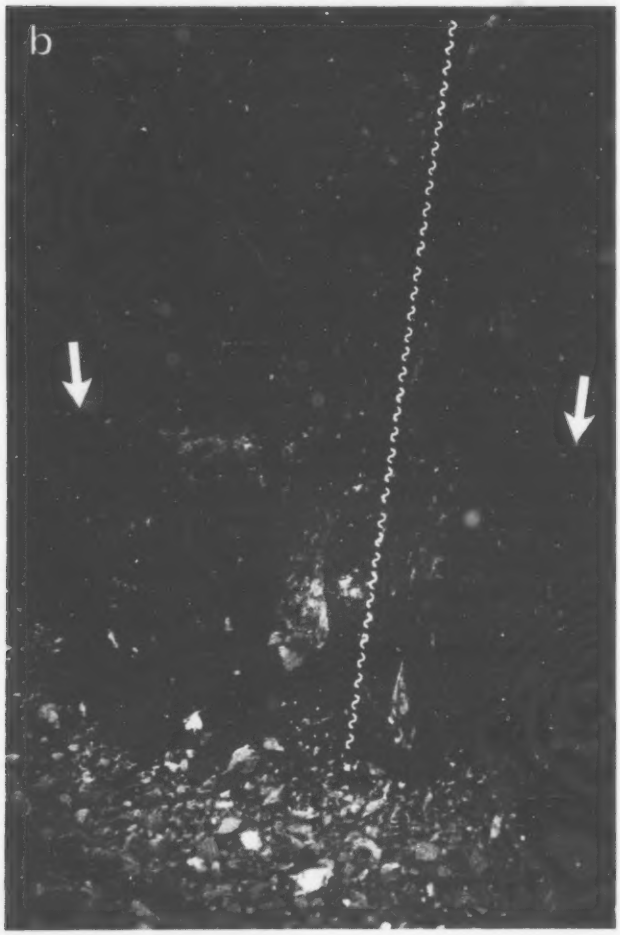


Figure GS-15-8: Shallowly southeast-dipping (flat) quartz-carbonate-sulphide vein. Note the sulphidic halo developed where the vein intersects iron–formation, and the slight ductile bending of the bedding lamellae adjacent to the vein.

Figure GS-15-9: a) Emplacement of quartz-sulphide vein (QSV) following F_4 folding (arrow); b) QSV 'flat' emplacement following faulting (arrow).



to 60 cm wide) of intense sulphidization of the host iron-formation. In the East pit, the sulphidization is manifested as replacement of the magnetite lamellae by pyrite (and gold), preserving the primary laminations and secondary foliations. The sulphidization halos are developed around both the steeps and the flats. The link between emplacement of the mineralization halos and sulphidization of magnetite is demonstrated by the lack of halo development where the QSV cut the interbedded argillite or intermediate dykes.

The low-temperature alteration assemblages in the later fabrics in all phases of the intrusions indicate that gold mineralization entirely postdates all intrusive events. As such, the gold is not associated with magmatic-hydrothermal activity related to these intrusive phases. It is possible that the late fabrics formed preferentially along the intrusion's north margin, which is the southern margin of the iron-formation–hosted gold mineralization and the focus of gold-mineralizing hydrothermal-fluid flow in the region; this could be due to the competency contrast between the iron-formation and the intrusion. Although this is consistent with the orientation of the southwest-dipping mineralized flats (faults/shears) and S_2 parallel steeps (phyllonite zones) in the area, the intermittent gold mineralization along the intrusive contact is not entirely consistent with this scenario, although the obvious control of sulphidization processes by magnetite may explain the minimal mineralization evident in the intermediate intrusion.

CONCLUSIONS

The intermediate Pool Lake suite intrusion is a zoned or composite intrusion with a massive granodioritic core grading to a marginal phase of dioritic composition. Although difficult to ascertain due to the variability of alteration, the granodioritic and tonalitic phases have a metaluminous volcanic-arc (I-type) affinity, whereas the diorite seems to have an Fe-rich tholeiitic character, possibly related to a first-phase extensional event associated with intrusion emplacement. The increase in fabric development and at least two stages of hydrothermal alteration (high T and low T) are recognized petrographically and seem to characterize the zone. An earlier foliation (S_2) in the outer margin, coupled with the generally discordant nature of the intrusion, suggests that it was emplaced syn- to post-F_2 folding and fabric development, which is consistent with the peak metamorphic mineralogy. Analogous to the late controls on gold-sulphide-quartz veining in the iron-formation–hosted sequence, the late chlorite-quartz-sulphide veins that locally contain gold in the intrusion are associated with faults and shears that seem to exploit the earlier S_2 fabric and related structures. As outlined before and based on fabric overprinting relationships, the gold mineralization is controlled by late brittle shears/joints that parallel the steep S_1-S_2 fabrics (steeps) and shallowly southwest-dipping brittle-ductile faults (flats) in the iron-formation and argillite. It is possible, especially considering the intensity of late shearing and low-temperature alteration along the intrusion-supracrustal contact, that the faulting and shearing were controlled, in part, by the competency contrast between the iron-formation and the altered intrusion margin. Although it does not seem that the intrusion played a direct role in the mineralizing process, it is probable that late auriferous hydrothermal solutions were focused from depth by these

structures and enhanced fabrics. Due to the selective nature of the sulphidization process (mainly of magnetite) that seems to be directly linked to mineralization, the intermediate intrusion was a less favourable host to the mineralization.

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MINERALIZATION-POTENTIAL MAPPING: A DATA-FUSION ANALYSIS

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SUMMARY

This project is a collaboration between the University of Manitoba and the Manitoba Geological Survey, with the objective of creating a regional database of geological and geophysical data for the Lynn Lake greenstone belt, and producing mineralization-potential maps using a GIS platform and data-fusion techniques. Data sets compiled include regional geology, lithogeochemistry, lake sediment and till geochemistry, airborne geophysics and synthetic aperture radar (SAR) data. A preliminary set of mineralization-potential maps for shear-hosted and VMS-type deposits has been completed using regional geology, lithogeochemistry and geophysical data. Spatial deposit models were developed for the two deposit types and fuzzy-logic data-fusion methods were used to integrate map layers. Preliminary results show good correlation between the potential maps, produced using theoretical principles, and known mineral occurrences throughout the southern portion of the Lynn Lake greenstone belt.

INTRODUCTION

Geographic information systems (GIS) have been developed over the years to integrate and manage large sets of diverse spatial data. These systems are also capable of manipulating the data to analyze and model the interrelations between spatial data sets (Bonham-Carter, 1994). The manipulation, or integration, of the spatial data can be accomplished using statistical weighting methods, Boolean logic, fuzzy logic, Dempster-Shafer theory, and neural-network data-fusion methods. Combinations of data can be used to locate areas of potential economic interest, define regional bedrock geology, or delineate major structural trends. However, data combination or fusion must be done in a consistent and justifiable manor; otherwise, the results will not be precise and technically accurate (Argialas and Harlow, 1990). Another important feature of this approach is that one must also understand that the integration of data is concentrated on a two-dimensional surface, even though particular data sets, such as magnetic data, are defined by three-dimensional structures (Moon, 1993). Different spatial resolution among data sets, together with the incomplete and imprecise nature of spatial data sets, requires a proper quantitative method of integration to address such problems (Moon, 1993).

Geological and geophysical data are commonly represented in the form of spatial data because a vast amount of information can be shown and understood far more easily than if the same data were presented in a written format. When spatial data sets are used together, the amount of information that can be acquired increases significantly; however, the method of integration used can give substantially different results. In the case of mineral-potential mapping, each layer of spatial data must be weighted according to the particular type of deposit being explored. It is here where simple combination does not work and the introduction of theoretical principles of geology and geophysics play a major role. Determining the various spatial data sets to be used, and the method of data fusion, is controlled by a mineral deposit model that, in turn, also determines the weighting of spatial data layers. In areas where numerous mineral occurrences are known, statistical data-fusion methods are useful. In many areas, however, previous exploration is minimal, so subjective methods are more applicable. In the case of minimal exploration, a geologist has complete control over the integration approach and weighting of the spatial data. This type of integration does not have the same quantitative accuracy as statistical methods. Subjective methods do, however, allow for flexibility, thus giving the geologist the ability to test many different scenarios, or spatial deposit models. These same methods can also be applied to less specific targets, such as regional-scale structures, by integrating aeromagnetic and SAR data.

PROJECT OBJECTIVE AND BASIC METHODOLOGY

The objective of this project is to produce structural and mineral-potential maps for the southern part of the Lynn Lake greenstone belt, specifically with respect to the Johnson Shear Zone (JSZ) and associated structures. This will be accomplished by using the theoretical principles of geology and geophysics to develop a conceptual mineral deposit model, which is then applied to a set of spatial data using fuzzy logic and Dempster-Shafer data-fusion methods. A potential map of major structural trends will be completed first. The structure map will, in turn, be applied to a number of spatial deposit models for mineralization-potential mapping.

Framework for Basic Methodology (Bonham-Carter, 1994)

1) Establish a conceptual mineral deposit model: A well defined model is key when applying theoretical principles of geology and geophysics to develop accurate mineral-potential maps. The two data-fusion methods that will be used have been developed for areas with minimal exploration. The weights applied to the various spatial data sets are not accomplished statistically, but are subjectively based on a spatial deposit model. Hence, a more defined model will result in more accurate mineral-potential maps.
2) Build a spatial database: Data sets include regional geological and structural maps, airborne geophysical data (very low frequency electromagnetics [VLF], total-field magnetics [TF], vertical-gradient magnetics [VG]), RADARSAT SAR data, regional lithogeochemistry, regional lake sediment geochemistry and regional till geochemistry. Note that the RADARSAT data, regional lake sediment geochemistry and regional till geochemistry were not integrated in the preliminary analysis.
3) Data analysis: Secondary spatial data sets will be developed to enhance or modify the usefulness of the existing spatial data sets by:
   - developing new maps based on pre-existing spatial data sets and other nonspatial data available with respect to the given mineral deposit model;
   - enhancing various maps and integrating to further define major structural features (specifically the JSZ), using the existing regional structural mapping, geophysical, and RADARSAT SAR data; and
   - producing proximity maps that can address both spatial associations and decay rates, linear or exponential, related to a particular spatial data set or group of data sets.
4) Integration of spatial data maps: This is accomplished using data-fusion methods and the given spatial deposit model.

MINERAL DEPOSIT MODEL

Known gold deposits within the southern portion of the Lynn Lake

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belt are characterized by intensive silicification and are associated with the JSZ. In order to locate new areas of economic interest, one must define a deposit model that best represents a particular target orebody. Deposit models can be quite simple and incorporate more that one specific target; however, a more defined model will produce a more accurate potential map. The model will also determine which spatial data sets should be incorporated, and how they should be used during integration. Many of the ideas defined by a deposit model cannot be treated in a spatial sense. Once a deposit model has been chosen, it must be utilized to represent the data available, so a spatial deposit model is developed. Nonspatial data, such as timing of structural and intrusive events, can also be incorporated into the spatial deposit model, but their influence is indirect and applied primarily in the weighting of the spatial data. Representation of a deposit model in a spatial sense is a major objective of this project. The spatial deposit models defined in this report are preliminary, and modifications are ongoing to develop more accurate potential maps. A detailed discussion of the major theoretical components related to the two deposit models will not be given here. For a more complete discussion, please refer to Roberts (1988), Bursnall (1989), Hodgson (1993) and Bonham-Carter et al. (1993) for the shear-hosted model; and Lydon (1988a, b), Franklin (1993) and Wright and Bonham-Carter (1996) for the VMS model.

Shear-Hosted Magmatic Hydrothermal Deposit Model

Thrust and strike-slip faults are common structural features associated with shear zones, commonly located at or close to the transition from brittle to ductile regions in the crust (between 7 and 12 km in depth). Within shear zones, a hydrous component typically extracts elements from adjacent wall rocks and transfers them to areas where they precipitate in greater concentration. Fluids liberated during magmatic solidification or during metamorphic dehydration usually make up the hydrous component. Spatial association to source rocks is common, although zones of mineralization are not necessarily specific to a particular rock type. Some rocks tend to be more susceptible to alteration or fracturing, and are therefore more probable sites for element concentration. The key components of this model then become, with respect to degree of decreasing importance, structural control, source of the hydrous component, and source of the mineral component.

Volcanic-Hosted Massive Sulphide (VMS) Deposit Model

Hydrothermal fluids circulating through a package of subaqueous mafic and felsic volcanic rocks is the basic environment for the formation of VMS deposits. The key components for this model include a heat source, which drives the hydrothermal system, a fluid source and an impermeable cap to the hydrothermal reaction zone. High fluid temperatures are required to dissolve a sufficient amount of metals for later deposition. The source of the heat is usually attributed to intrusive complexes and associated dykes and sills. This heat source must be great enough to overcome cool seawater, the major source of the fluids, circulating through the permeable subaqueous volcanic rocks. Magmatic fluids are considered to be a minor component. Precipitated metals associated with the subaqueous volcanic rocks are sealed in with an impermeable cap, commonly massive pyroclastic ash or rhyolitic flows. Faulting within the region may act as a trap for the precipitating sulphide minerals, although structural control is a minor component in the formation of a VMS deposit. Alteration is a key spatial component in both VMS and shear-hosted deposits; however, a complete spatial data set is not available, so alteration is not discussed here.

SPATIAL DATA SET

Development of a spatial data set will be accomplished using GIS and will include spatial and point data. Data sets acquired for this preliminary study include regional geology maps, lithogeochemistry and regional geophysics (Table GS-16-1). The extent of the individual data sets varies and they do not cover the entire greenstone belt; however, the problems associated with missing or incomplete data will be addressed during map integration.

Table GS-16-1: Spatial data sets used in preliminary mineralization-potential maps.

1:50 000 scale geological map, Lynn Lake greenstone belt; Manitoba Geological Survey.
1:250 000 scale geochemical units, Lynn Lake greenstone belt; Manitoba Geological Survey, OF99-13 (Zwanzig et al., 1999).
Airborne TF, VG, VLF-EM, Project 10500, Barrington Lake; Canada–Manitoba joint agreement, 1983.
Airborne TF, VG, VLF-EM, Project 13800, Lynn Lake; Canada–Manitoba joint agreement, 1982.
Airborne TF, VG, VLF-EM, Project 18201, Barrington Lake; Canada-Manitoba Mineral Development Agreement (MDA), 1985.

DATA ANALYSIS

Shear-Hosted Model

Structural control is the major component of the shear-hosted model. Therefore, a structure (lineament) map has been produced by combining known major fault and shear zone features, compiled during regional mapping, and geophysical data. Synthetic aperture radar (SAR) data have been acquired for this project, but have not yet been integrated into the structure mapping. Lineaments were defined from the vertical-gradient magnetic map by applying directional and gradient filters, and by adjusting the image threshold value. Lineaments were divided into four categories based on their length. Proximity maps, at 100 m intervals to a maximum of 5 km, were created for each of the lineament maps and the known structure map.

Heat can be attributed to intrusive units, whereas the fluid sources can be related to fluid loss during magmatic crystallization. 'Proximity to heat source' maps were created by selecting intrusive units from the geological map. Proximity maps were created for both mafic and felsic intrusive units with 100 m intervals to a maximum of 5 km.

Reclassification, or map recoding, is the reassignment of the classes of a spatial data layer to produce a new map. Reclassification is not used to change spatial relationships within a particular map layer, but instead to simplify or make integration between maps possible. Geology has been reclassified to combine similar lithological units. It has also been reclassified to represent relative rheological characteristics of intrusive, volcanic and sedimentary units. The lithogeochemical map has been reclassified into two maps, 'MgO content' and 'primitive versus evolved', based on the average MgO and rare-earth element (REE) data, respectively, for each of the defined geochemical units.

VMS Spatial Model

'Proximity to heat source' maps for the VMS spatial model were derived from the shear-hosted model 'proximity to heat source' maps; however, weighting of the maps is different to reflect the given model. Structure is a minor component for this model, so lineament maps and known structure have been combined into a single 'proximity to structure' map. Additional stratigraphic, or 'proximity to volcanic rocks', maps have been created for the VMS model using the geological map. The lithogeochemical data for volcanic units have not been applied in this preliminary analysis.

MAP INTEGRATION

The integration of the various data sets in this preliminary analysis has made use of the fuzzy-logic data-fusion method. The fuzzy-logic method utilizes the continuous scale from 0 to 1, or fuzzy membership function, which represents the full nonmembership (0) and full membership (1) function, respectively, of a fuzzy set. The initial advantages of the fuzzy-logic method are its straightforward implementation and its ability to develop mineralization-potential maps in areas that are not well explored. Fuzzy-set theory enables the user to develop a set of attributes for a particular spatial data set, whereby membership functions can be expressed as linear or nonlinear, depending on the problem at hand (Bonham-Carter, 1994). Defining a fuzzy set, A in X, as the set of ordered pairs

$$A = \{ x , u(x)\}$$

where x is a member of X and u(x) is the membership function, we can now develop a series of fuzzy sets for our geology, geochemistry, 'proximity to heat source', 'proximity to volcanic rocks', 'proximity to structure' and geophysical spatial data.

Five methods of map integration, fuzzy AND, fuzzy OR, fuzzy algebraic product, fuzzy algebraic sum and the gamma operation, have been defined for the fuzzy-logic method (Bonham-Carter, 1994). The first two methods are not used in this project, and the algebraic product and sum methods are only used in combination with the gamma operation. The algebraic product is defined as

$$\text{Ucombination } (x_1, x_2, \dots , x_n) = (x_1)\,(x_2) \dots (x_n)$$

Its output is less than or equal to the smallest contributing membership value and therefore represents the minimum probability. The algebraic sum, on the other hand, results in a maximum probability that is equal to or greater than the largest contributing factor:

$$\text{Ucombination } (x_1, x_2, \dots , x_n) = 1 - \{(1 - (x_1))\,(1 - (x_2)) \dots (1 - (x_n))\}$$

The last method is the gamma operation, where a parameter, gamma (γ), is used to ensure a flexible compromise between the maximum probability of the algebraic sum and the minimum probability of the algebraic product (Bonham-Carter, 1994). The value of γ can range between 0 and 1, where 0 would result in the output being equal to the algebraic product, and 1 would result in the output being equal to the algebraic sum. With a γ value of 0.75 used in this project, the gamma operation can be defined as:

$$\text{Ucombination } (x_1, x_2, \dots , x_n) = ((\text{fuzzy algebraic sum})^{\gamma})\,((\text{fuzzy algebraic product})^{1-\gamma})$$

Before map integration can take place, weighting of the various maps must be completed. Weighting is completed at two levels: first between the individual maps, and second with regard to classes within the individual maps themselves. Tables GS-16-2 and -3 list the fuzzy-set membership functions for each of the given spatial data sets for the shear-hosted and VMS deposit models, respectively. Between individual maps, a maximum possible membership value is given to a particular map based on the degree of evidence of that map. Weighting of the various classes for each map has been applied differently with respect to the lithological unit map, geochemical maps, geophysical maps and proximity maps. The lithological and 'primitive versus evolved' maps are subjectively weighted. The 'MgO map' is based on the relative MgO values for the given volcanic units. Geophysical maps are reclassified relative to the original image data values. 'Proximity to heat' and 'proximity to structure' maps are weighted with an exponential decay function from proximal to distal, whereas a linear decay function is applied to the 'proximity to volcanic rocks' maps.

Figure GS-16-1 is an integration flow chart representing the preliminary shear-hosted spatial deposit model. The model is separated into two major components, the source zone probability and deposition zone probability. The source zone probability relates to the source of the mineral component and is dependent on primary lithology and geochemistry, whereas the deposition zone probability is dependent on structural control and source of the hydrous component, primarily magmatic in this case. Structural control is directly related to the rheology of a particular rock type and the spatial associations between rock types. Geophysical data represent present geophysical characteristics of the regional geology. Therefore, both rheology and geophysics have tentatively been included in the depositional zone probability.

Figure GS-16-2 is the integration flow chart for the VMS spatial deposit model. This model has been separated into source zone probability, geophysical factor and deposition zone probability. The geophysical data again represent present geophysical characteristics of the regional geology; however, because of the magnetic nature of VMS deposits, the geophysical factor is considered as a separate component.

Figures GS-16-3 and -4 are the resulting mineralization-potential maps produced for the two deposit models, where the grey-scale range is relative probability. Only the top 20th percentile is shown, with grey-scale values being further stretched to highlight local variations. Three geophysical surveys were completed in the area, but they have not been levelled relative to one another. Therefore, each mineralization-potential map comprises three individual maps, each with its own relative potential scale that represents the special limits of the particular geophysical survey.

PRELIMINARY RESULTS

Preliminary spatial deposit models have been completed and the resulting maps were used during the 2000 summer field season. Shear-hosted and VMS deposit models were integrated using the fuzzy-logic data-fusion method. Initial analysis shows that high-potential sites, specifically within the top 20th potential, show good agreement with known mineralization in this part of the Lynn Lake greenstone belt. It must be stressed here that these mineralization-potential maps are based on spatial deposit models defined by theoretical principles of geology and geophysics, and are not defined by known occurrences within the region. The initial analysis was based on known occurrences compiled in the Mineral Deposit Series and from ground-truthing completed during the 2000 field season. A detailed analysis of the preliminary and subsequent maps will be completed over the winter (2000–2001) and will include further ground-truthing. Analytical data from geochemical and assay samples collected during the 2000 field season have not yet been received, but will be included in the subsequent analysis. Shear-zone potential maps were completed for vein-type and disseminated-type occurrences. The two shear-hosted potential maps are similar (disseminated-type potential map is shown in Fig. GS-16-3), suggesting that either rheology plays a minor role or that the application of rheology is incorrect. Further work is necessary to address this issue. A number of high-potential zones have been located where there has been no known major exploration. These areas, if truly unexplored, could represent new areas of interest and would be the focus of subsequent work.

One of the major problems of integration when using the fuzzy-logic data-fusion method is areas of no data, because all subsequent integration must be multiplied by zero. This problem was overcome by giving areas of no data a non-zero value of 0.01. The results showed that, even with the non-zero value, the method does not represent these areas sufficiently. Subsequent analysis will make use of the Dempster-Shafer method, which can represent missing or zero data values.

Table GS-16-2: Fuzzy set membership function for shear-hosted spatial deposit model.

Metres	PXL5	PXL2	PXL1	PXL0	PXHF	PXHM
0-100	0.80	0.60	0.50	0.40	0.60	0.40
100-200	0.77	0.58	0.48	0.38	0.58	0.38
200-300	0.74	0.55	0.46	0.37	0.55	0.37
300-400	0.71	0.53	0.44	0.35	0.53	0.35
400-500	0.68	0.51	0.42	0.34	0.51	0.34
500-600	0.65	0.49	0.41	0.32	0.49	0.32
600-700	0.62	0.46	0.39	0.31	0.46	0.31
700-800	0.59	0.44	0.37	0.30	0.44	0.30
800-900	0.56	0.42	0.35	0.28	0.42	0.28
900-1000	0.54	0.40	0.34	0.27	0.40	0.27
1000-1100	0.51	0.38	0.32	0.26	0.38	0.26
1100-1200	0.49	0.37	0.30	0.24	0.37	0.24
1200-1300	0.46	0.35	0.29	0.23	0.35	0.23
1300-1400	0.44	0.33	0.27	0.22	0.33	0.22
1400-1500	0.41	0.31	0.26	0.21	0.31	0.21
1500-1600	0.39	0.29	0.25	0.20	0.29	0.20
1600-1700	0.37	0.28	0.23	0.18	0.28	0.18
1700-1800	0.35	0.26	0.22	0.17	0.26	0.17
1800-1900	0.33	0.25	0.20	0.16	0.25	0.16
1900-2000	0.31	0.23	0.19	0.15	0.23	0.15
2000-2100	0.29	0.22	0.18	0.14	0.22	0.14
2100-2200	0.27	0.20	0.17	0.13	0.20	0.13
2200-2300	0.25	0.19	0.16	0.13	0.19	0.13
2300-2400	0.23	0.17	0.15	0.12	0.17	0.12
2400-2500	0.22	0.16	0.14	0.11	0.16	0.11
2500-2600	0.20	0.15	0.13	0.10	0.15	0.10
2600-2700	0.18	0.14	0.12	0.09	0.14	0.09
2700-2800	0.17	0.13	0.11	0.08	0.13	0.08
2800-2900	0.15	0.12	0.10	0.08	0.12	0.08
2900-3000	0.14	0.11	0.09	0.07	0.11	0.07
3000-3100	0.13	0.10	0.08	0.06	0.10	0.06
3100-3200	0.12	0.09	0.07	0.06	0.09	0.06
3200-3300	0.10	0.08	0.06	0.05	0.08	0.05
3300-3400	0.09	0.07	0.06	0.05	0.07	0.05
3400-3500	0.08	0.06	0.05	0.04	0.06	0.04
3500-3600	0.07	0.05	0.05	0.04	0.05	0.04
3600-3700	0.06	0.05	0.04	0.03	0.05	0.03
3700-3800	0.05	0.04	0.03	0.03	0.04	0.03
3800-3900	0.05	0.03	0.03	0.02	0.03	0.02
3900-4000	0.04	0.03	0.02	0.02	0.03	0.02
4100-4200	0.03	0.02	0.02	0.02	0.02	0.02
4200-4300	0.03	0.02	0.02	0.01	0.02	0.01
4300-4400	0.02	0.02	0.01	0.01	0.02	0.01
4400-4500	0.02	0.01	0.01	0.01	0.01	0.01
4500-4600	0.01	0.01	0.01	0.01	0.01	0.01
4600-4700	0.01	0.01	0.01	0.01	0.01	0.01
4700-4800	0.01	0.01	0.01	0.01	0.01	0.01
4800-4900	0.01	0.01	0.01	0.01	0.01	0.01
4900-5000	0.01	0.01	0.01	0.01	0.01	0.01
5000+	0.01	0.01	0.01	0.01	0.01	0.01

TF	VLF
0.40	0.40
0.39	0.39
0.38	0.38
0.38	0.38
0.37	0.37
0.36	0.36
0.35	0.35
0.34	0.34
0.34	0.34
0.33	0.33
0.32	0.32
0.31	0.31
0.30	0.30
0.30	0.30
0.29	0.29
0.28	0.28
0.27	0.27
0.26	0.26
0.26	0.26
0.25	0.25
0.24	0.24
0.23	0.23
0.22	0.22
0.22	0.22
0.21	0.21
0.20	0.20
0.19	0.19
0.18	0.18
0.18	0.18
0.17	0.17
0.16	0.16
0.15	0.15
0.14	0.14
0.14	0.14
0.13	0.13
0.12	0.12
0.11	0.11
0.10	0.10
0.10	0.10
0.09	0.09
0.08	0.08
0.07	0.07
0.06	0.06
0.06	0.06
0.05	0.05
0.04	0.04
0.03	0.03
0.02	0.02
0.02	0.02
0.01	0.01

Unit	Fuzzy
1	0.00
2	0.60
3	0.60
4	0.50
5	0.40
6	0.20
7	0.20
8	0.05
9	0.10
10	0.01
11	0.01
12	0.01
13	0.01
14	0.45
15	0.45
16	0.45
17	0.10
18	0.10
19	0.45
20	0.10
21	0.10
22	0.10
23	0.10

G-unit	f_{MgO}	f_{prim}
1a	0.25	0.30
1b	0.36	0.30
1c	0.29	0.30
1d	0.21	0.30
2a	0.60	0.10
2b	0.58	0.10
2c	0.35	0.10
2d	0.19	0.10
3a	0.35	0.25
3b	0.16	0.25
3c	0.23	0.25
3d	0.21	0.25
4a	0.24	0.25
4b	0.23	0.25
4c	0.22	0.25
5a	0.56	0.01
6a	0.28	0.35
6b	0.00	0.01
6c	0.21	0.35
7a	0.32	0.50
7b	0.13	0.50
7c	0.33	0.50
8a	0.42	0.60
8b	0.26	0.60
9a	0.73	0.01
10a	0.06	0.01
10b	0.08	0.01
11a	0.00	0.01
11b	0.03	0.01
11c	0.06	0.01
11d	0.04	0.01
11e	0.03	0.01
11f	0.01	0.01
11g	0.01	0.01
12	0.01	0.01
13	0.01	0.01
14	0.01	0.01
15	0.01	0.01
16	0.01	0.01
17	0.01	0.01

Unit	R_{vein}	R_{diss}
1	0.00	0.00
2	0.10	0.10
3	0.10	0.10
4	0.10	0.10
5	0.10	0.10
6	0.10	0.10
7	0.10	0.10
8	0.05	2.00
9	0.05	2.00
10	0.05	2.00
11	0.05	2.00
12	0.05	2.00
13	0.05	2.00
14	0.20	0.05
15	0.20	0.05
16	0.20	0.05
17	0.20	0.05
18	0.20	0.05
19	0.20	0.05
20	0.20	0.05
21	0.20	0.05
22	0.20	0.05
23	0.20	0.05

Explanation of abbreviations:

PXL5, proximity to lineaments ≥ 5 km in length

PXL2, proximity to lineaments 2–5 km in length

PXL1, proximity to lineaments 1–2 km in length

PXL0, proximity to lineaments 0.5–1 km in length

PXHF, proximity to heat, felsic intrusive units

PXHM, proximity to heat, mafic intrusive units

TF, total-field magnetics

VLF, very low frequency electromagnetics

Unit (lithological units): 1, undivided; 2, aphyric basalt; 3, porphyritic basalt; 4, mafic and intermediate volcanic rocks, amphibolite; 5, intermediate and felsic volcanic rocks; 6, dacite; 7, rhyolite; 8, conglomerate; 9, sedimentary rocks, paragneiss; 10, conglomerate, greywacke, 11, Burntwood Group greywacke, siltstone, mudstone; 12, Sickle Group conglomerate; 13, Sickle Group sandstone; 14, gabbro, diabase; 15, hornblende diorite, quartz diorite; 16, gabbro, norite, diorite, ultramafic rock; 17, granodiorite-diorite; 18, granite-granodiorite; 19, mafic intrusions; 20, granodiorite-tonalite; 21, granite-granodiorite; 22, granitoid rock; 23, quartz porphyry

Fuzzy, fuzzy membership value

G-unit (geochemical units): 1, arc tholeiite; 2, calc-alkaline basalt to andesite; 3, enriched arc tholeiite; 4, depleted arc tholeiite; 5, komatiitic basalt; 6, weakly depleted arc tholeiite; 7, arc tholeiite transitional to E-MORB; 8, MORB-like basalt; 9, ocean-island basalt (OIB)-ultramafic rock; 10, dacite; 11, rhyolite; 12, volcanogenic sedimentary rocks; 13, Zed Lake greywacke; 14, Ralph Lake conglomerate; 15, Burntwood Group (1.86–1.84Ga); 16, Sickle Group; 17, intrusive rocks (1.82–1.96Ga)

f_{MgO}, fuzzy membership value for 'MgO'

f_{Prim}, fuzzy membership for 'primitive versus evolved'

R_{Vein}, rheology vein type

R_{diss}, rheology disseminated type

Table GS-16-3: Fuzzy set membership function for VMS spatial deposit model.

Metres	PXL	PXHF	PXHM
0-100	0.20	0.40	0.40
100-200	0.19	0.38	0.38
200-300	0.18	0.37	0.37
300-400	0.18	0.35	0.35
400-500	0.17	0.34	0.34
500-600	0.16	0.32	0.32
600-700	0.15	0.31	0.31
700-800	0.15	0.30	0.30
800-900	0.14	0.28	0.28
900-1000	0.13	0.27	0.27
1000-1100	0.13	0.26	0.26
1100-1200	0.12	0.24	0.24
1200-1300	0.12	0.23	0.23
1300-1400	0.11	0.22	0.22
1400-1500	0.10	0.21	0.21
1500-1600	0.10	0.20	0.20
1600-1700	0.09	0.18	0.18
1700-1800	0.09	0.17	0.17
1800-1900	0.08	0.16	0.16
1900-2000	0.08	0.15	0.15
2000-2100	0.07	0.14	0.14
2100-2200	0.07	0.13	0.13
2200-2300	0.06	0.13	0.13
2300-2400	0.06	0.12	0.12
2400-2500	0.05	0.11	0.11
2500-2600	0.05	0.10	0.10
2600-2700	0.05	0.09	0.09
2700-2800	0.04	0.08	0.08
2800-2900	0.04	0.08	0.08
2900-3000	0.04	0.07	0.07
3000-3100	0.03	0.06	0.06
3100-3200	0.03	0.06	0.06
3200-3300	0.03	0.05	0.05
3300-3400	0.02	0.05	0.05
3400-3500	0.02	0.04	0.04
3500-3600	0.02	0.04	0.04
3600-3700	0.02	0.03	0.03
3700-3800	0.01	0.03	0.03
3800-3900	0.01	0.02	0.02
3900-4000	0.01	0.02	0.02
4100-4200	0.01	0.02	0.02
4200-4300	0.01	0.01	0.01
4300-4400	0.01	0.01	0.01
4400-4500	0.01	0.01	0.01
4500-4600	0.01	0.01	0.01
4600-4700	0.01	0.01	0.01
4700-4800	0.01	0.01	0.01
4800-4900	0.01	0.01	0.01
4900-5000	0.01	0.01	0.01
5000+	0.01	0.01	0.01

Metres	PXR	PXIF
0-100	0.80	0.70
100-200	0.72	0.63
200-300	0.65	0.57
300-400	0.58	0.51
400-500	0.51	0.45
500-600	0.45	0.39
600-700	0.39	0.34
700-800	0.34	0.30
800-900	0.29	0.25
900-1000	0.24	0.21
1000-1100	0.20	0.18
1100-1200	0.16	0.14
1200-1300	0.13	0.11
1300-1400	0.10	0.09
1400-1500	0.07	0.06
1500-1600	0.05	0.04
1600-1700	0.03	0.03
1700-1800	0.02	0.02
1800-1900	0.01	0.01
1900-2000	0.01	0.01
2000+	0.01	0.01

Unit	Fuzzy
2	0.70
3	0.70
4	0.65
5	0.60
6	0.55
7	0.55
8	0.20
9	0.40
10	0.20
11	0.20
12	0.20
13	0.20
14	0.25
15	0.25
16	0.25
17	0.20
18	0.20
19	0.25
20	0.20
21	0.20
22	0.20
23	0.20

TF	VLF
0.50	0.70
0.49	0.69
0.48	0.67
0.47	0.66
0.46	0.64
0.45	0.63
0.44	0.62
0.43	0.60
0.42	0.59
0.41	0.57
0.40	0.56
0.39	0.55
0.38	0.53
0.37	0.52
0.36	0.50
0.35	0.49
0.34	0.48
0.33	0.46
0.32	0.45
0.31	0.43
0.30	0.42
0.29	0.41
0.28	0.39
0.27	0.38
0.26	0.36
0.25	0.35
0.24	0.34
0.23	0.32
0.22	0.31
0.21	0.29
0.20	0.28
0.19	0.27
0.18	0.25
0.17	0.24
0.16	0.22
0.15	0.21
0.14	0.20
0.13	0.18
0.12	0.17
0.11	0.15
0.10	0.14
0.09	0.13
0.08	0.11
0.07	0.10
0.06	0.08
0.05	0.07
0.04	0.06
0.03	0.04
0.02	0.03
0.01	0.01

Explanation of abbreviations:
PXL, proximity to lineaments 0.5–5 km in length
PXHF, proximity to heat, felsic intrusive units
PXHM, proximity to heat, mafic intrusive units
PXR, proximity to rhyolite
PXIF, proximity to intermediate and felsic volcanic rocks
Unit (lithological units): 1, undivided; 2, aphyric basalt; 3, porphyritic basalt; 4, mafic and intermediate volcanic rocks, amphibolite; 5, intermediate and felsic volcanic rocks; 6, dacite; 7, rhyolite; 8, conglomerate; 9, sedimentary rocks, paragneiss; 10, conglomerate, greywacke; 11, Burntwood Group greywacke, siltstone, mudstone; 12, Sickle Group conglomerate; 13, Sickle Group sandstone; 14, gabbro, diabase; 15, hornblende diorite, quartz diortite; 16, gabbro, norite, diorite, ultramafic rock; 17, granodiorite-diorite; 18, granite-granodiorite; 19, mafic intrusions; 20, granodiorite-tonalite; 21, granite-granodiorite; 22, granitoid rock; 23, quartz porphyry
Fuzzy, fuzzy membership value
TF, total-field magnetics
VLF, very low frequency electromagnetics

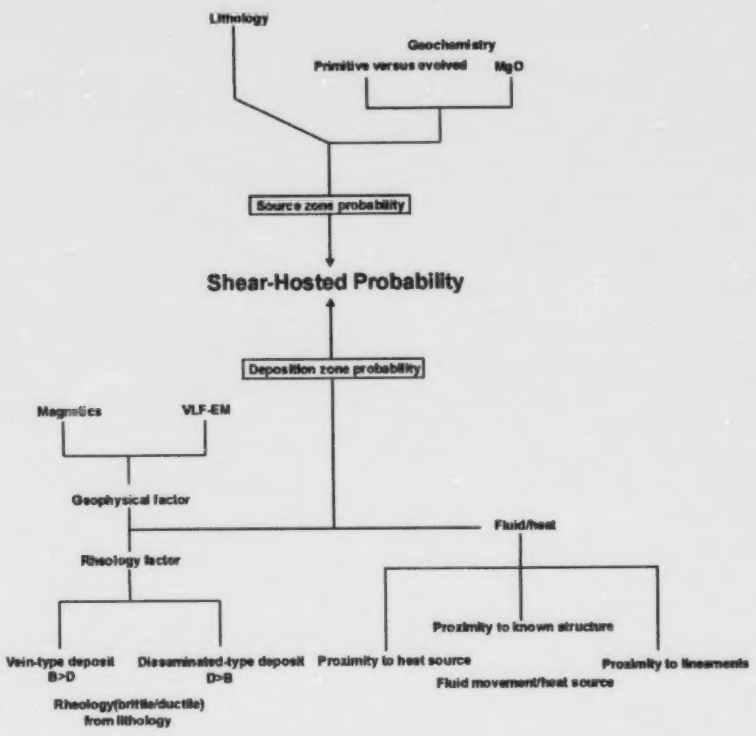

Figure GS-16-1: Integration flow chart for the shear-hosted magmatic hydrothermal spatial deposit model. The fuzzy-logic gamma operation is used at each stage of integration to minimize the increasive and decreasive effects of the algebraic product and algebraic sum.

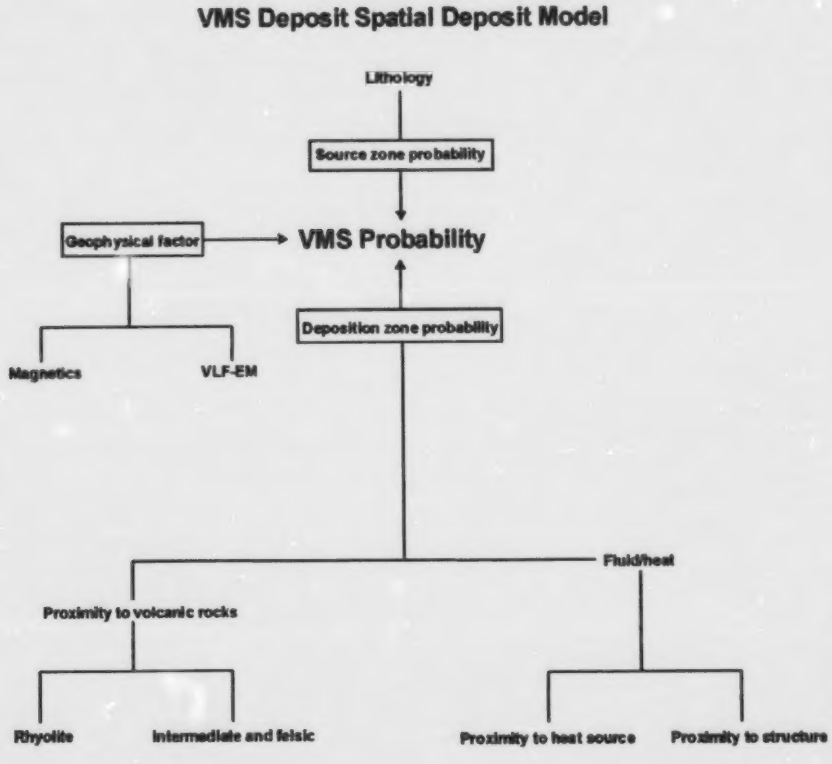

Figure GS-16-2: Integration flow chart for VMS spatial deposit model. The fuzzy-logic gamma operation is used at each stage of integration to minimize the increasive and decreasive effects of the algebraic product and algebraic sum.

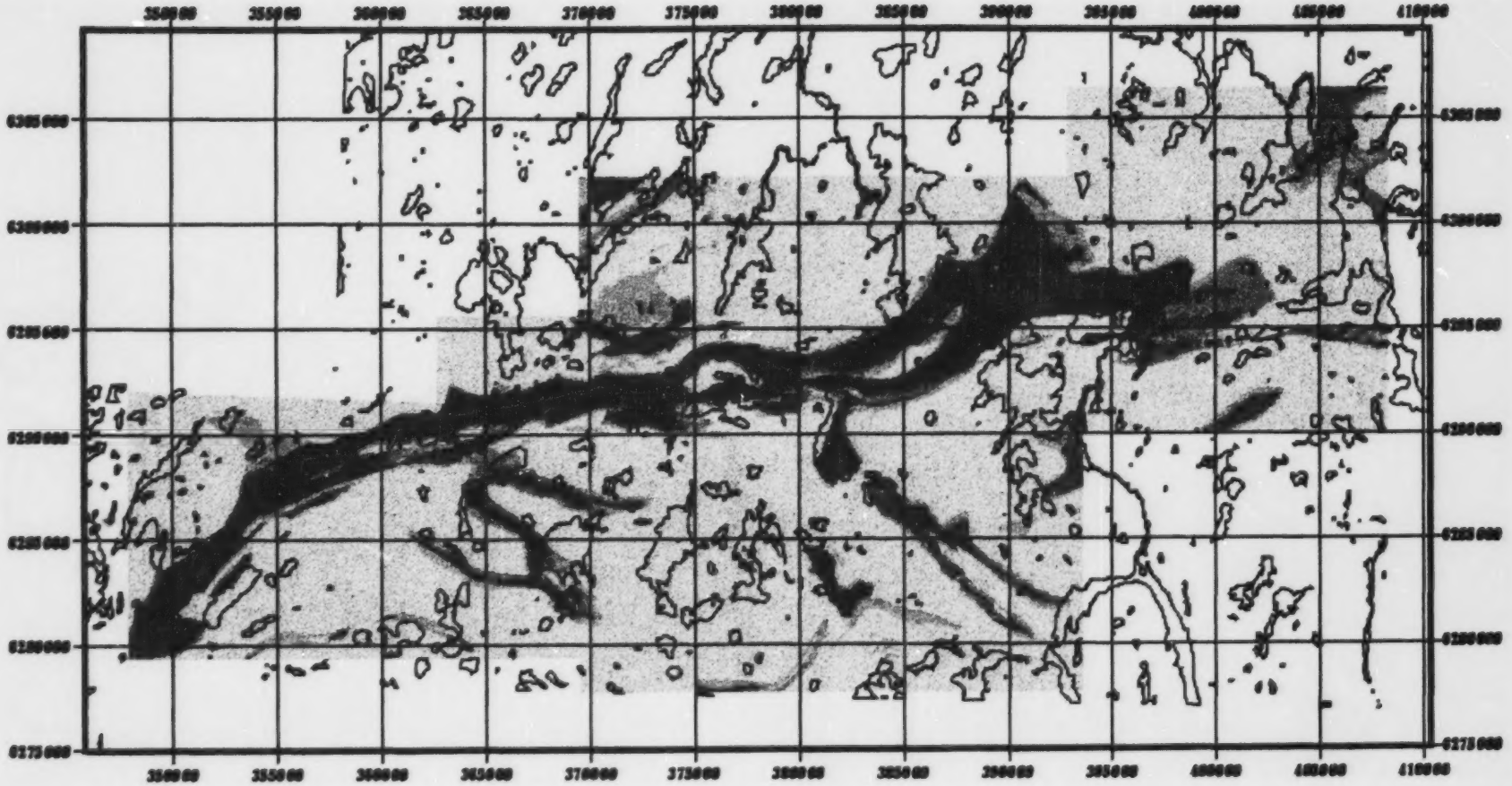

Figure GS-16-3: Mineralization-potential map for shear-hosted magmatic hydrothermal spatial deposit model. Grey-scale levels represent only the top 20th percentile, with relative scale being linear. White indicates low relative probability; black indicates high relative probability. Map base is UTM Zone 14, NAD83.

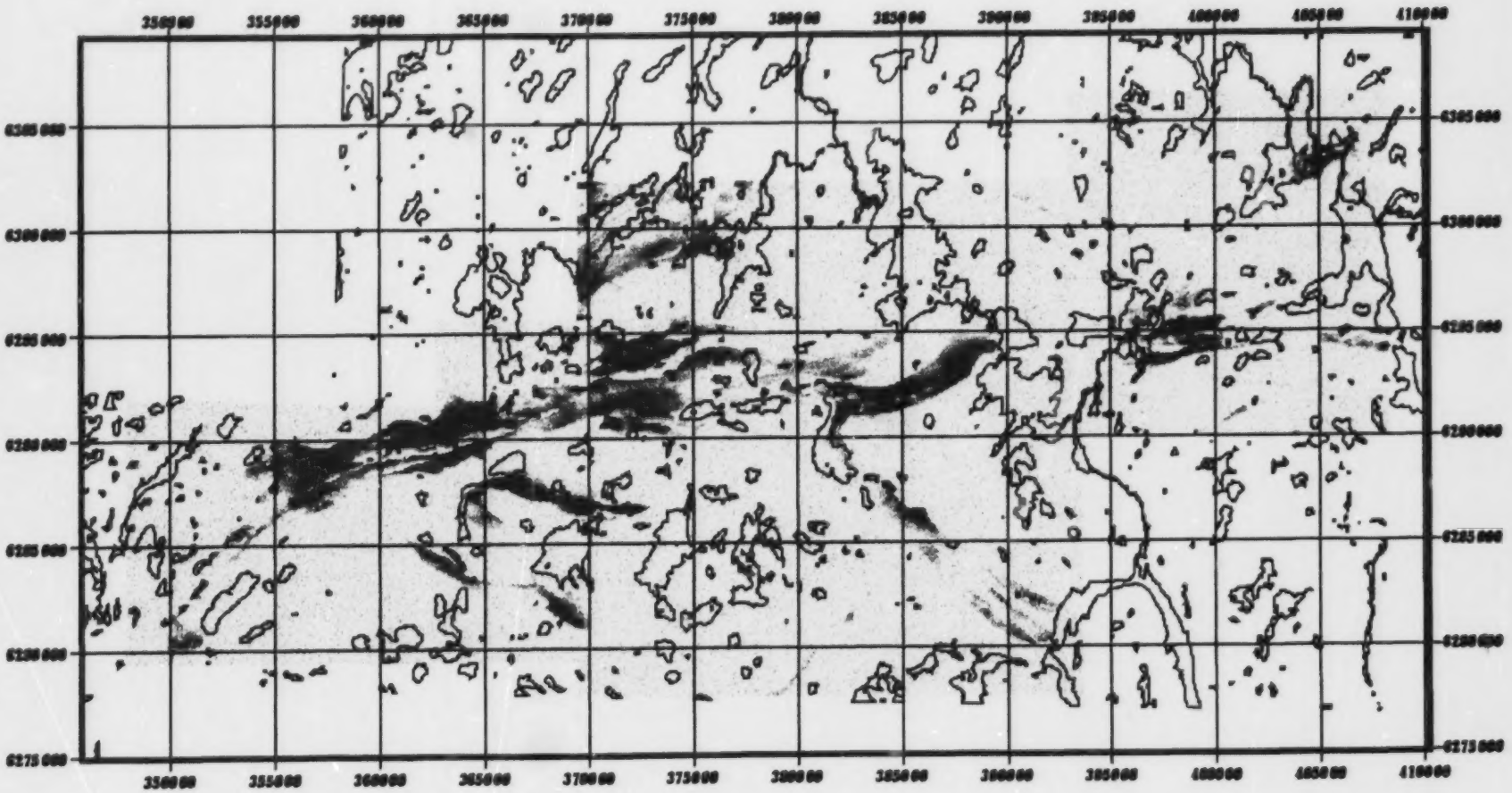

Figure GS-16-4: Mineralization-potential map for VMS spatial deposit model. Grey-scale levels represent only the top 20th percentile, with relative scale being linear. White indicates low relative probability; black indicates high relative probability. Map base is UTM Zone 14, NAD83.

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GS-17

GEOCHEMISTRY AND TECTONIC FRAMEWORK OF THE KISSEYNEW DOMAIN-LYNN LAKE BELT BOUNDARY (PART OF NTS 63P/13)

by H.V. Zwanzig

Zwanzig, H.V. 2000: Geochemistry and tectonic framework of the Kisseynew Domain-Lynn Lake belt boundary (part of NTS 63P/13); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 91-96.

SUMMARY

- Highly recrystallized, mafic to ultramafic volcanic rocks, previously mapped as undifferentiated amphibolite, fall into various geochemically distinctive units that were defined in better preserved rocks by Zwanzig et al. (1999).
- Despite alteration involving alkali elements and structural attenuation of the units, the composition of the least mobile elements is remarkably uniform for tens of kilometres along strike and provides a valuable tool for mapping.
- Assemblages of volcanic units have geochemical and metallogenic affinities to 1) volcanic arcs, 2) marginal basin floor (N-MORB and contaminated MORB), and 3) an ocean island.
- A complexly folded crustal suture between the turbidite units of the Kisseynew Domain and the arc volcanic rocks of the Lynn Lake belt contains the marginal-basin basalts.
- Major structural breaks separate the arc and marginal-basin assemblages, even where units were attenuated to a thickness of only a few metres; shear zones that reactivate or cut these breaks may host precious-metal deposits.

INTRODUCTION

Updated trace-element data have recently demonstrated that the Lynn Lake belt and the adjacent margin of the Kisseynew Domain contain numerous tectonostratigraphic assemblages (Zwanzig et al., 1999). Each assemblage represents a package of rocks with a distinctive stratigraphy, geochemistry and inferred plate-tectonic setting. Various assemblages were interpreted to have been structurally juxtaposed along major tectonic breaks. Whereas, in the Lynn Lake belt, the resulting structural stack comprises subvertical northeast- to southeast-trending panels, on the Kisseynew Domain margin, it consists of complexly refolded structural sheets. One unit (Laurie Lake assemblage) in a tightly folded sheet on the northeast shore of Laurie Lake was identified as the southwest arm of a prominent belt of calc-alkaline rocks (Zwanzig et al., 1999). Sheets to the southeast were found to include MORB-like marginal-basin basalts (Hatchet Lake basalt and Tod Lake basalt) and an ocean-island assemblage (Pickerel Narrows amphibolite). These rocks are unconformably overlain by conglomerate and arenite (Missi Group), which occur at the top of each structural sheet. The ocean-floor assemblages extend from Laurie Lake more than 100 km southeast to an area southeast of Granville Lake. They are a key component of the Granville Lake structural zone, which forms the north boundary of the Kisseynew Domain (Zwanzig, 1990). This zone has recently been interpreted as a crustal suture that developed during northward subduction and accretion of the MORB-like assemblages beneath the Lynn Lake belt, and subsequent underthrusting by Kisseynew turbidite basin fill during continental collision in the Trans-Hudson Orogen (White et al., 2000).

Specific units were re-examined in 2000 along the Kisseynew Domain boundary at the southwest end of the Lynn Lake belt (Fig. GS-17-1). Samples of volcanic rocks were collected and geochemical analyses were obtained from Activation Laboratories Ltd. (codes 4LITHO-MAJ ELEM FUS ICP and 4LITHO-TRACE ELM FUS

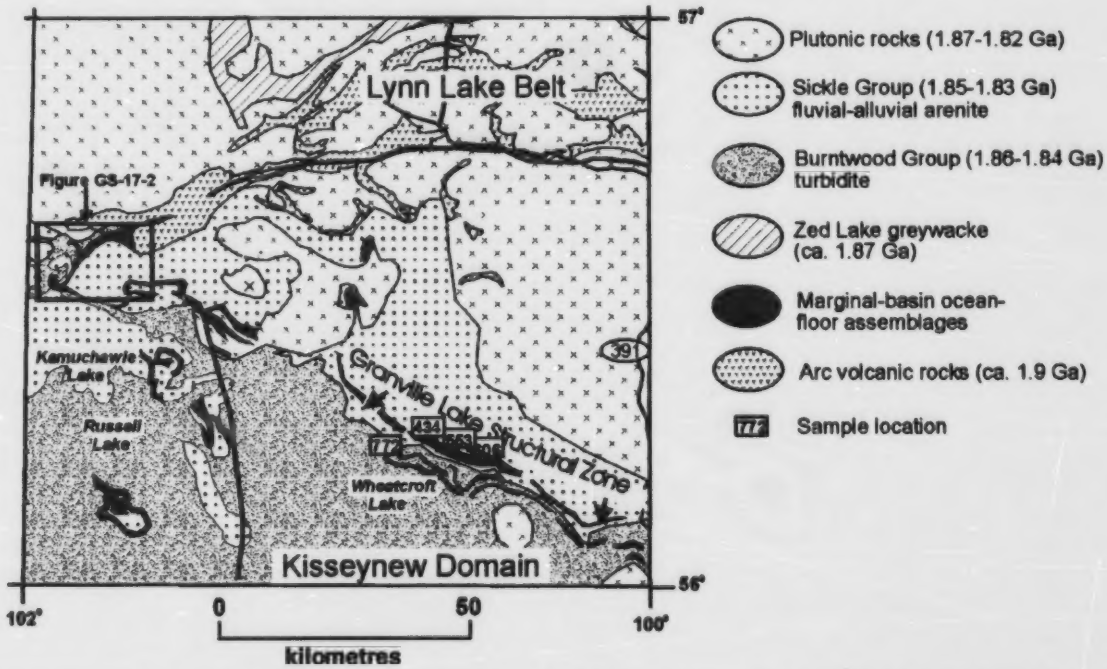

Figure GS-17-1: Geological map showing the Lynn Lake volcanic belt and Kisseynew Domain separated by the Granville Lake Structural Zone. Locations of selected geochemical samples are shown.

ICP/MS). Results will be used to help trace the units southwest across Laurie Lake and east along Eager Lake. Major- and trace-element geochemistry was also updated for existing samples from Granville Lake, Kamuchawie Lake and Russell Lake (Fig. GS-17-1). Preliminary results from 29 new analyses are summarized here and compared in simple variation diagrams to the data of Zwanzig et al. (1999). An important goal of this work is to delineate the Granville Lake suture zone at localities where it has only been inferred. Based on existing showings and a prominent As anomaly in glacial till to the south, this work may have application to precious-metal exploration.

A second objective is to trace probable splays of the Johnson Shear Zone along Tod Lake (Tod Lake Fault of Gilbert et al., 1980) and possibly across Laurie Lake (Fig. GS-17-2). To that end, a reconnaissance of shear zones was undertaken with C.J. Beaumont-Smith. Syn- to post-metamorphic (late collisional) shear fabrics, similar to those reported in Au-bearing zones in the central part of the Lynn Lake belt (Beaumont-Smith, GS-12, this volume), indicate that further work may provide promising results.

GEOCHEMICAL RESULTS

Mafic to ultramafic volcanic map units, each with a distinctive geochemical composition reported by Zwanzig et al. (1999), can be extended using the new data. These units have lost most of their primary structural features and were generally mapped as undifferentiated amphibolite in the new sampling area on Laurie and Eager lakes (Gilbert et al., 1980), as well as in parts of the Granville Lake area (Zwanzig, 1981), north of Kamuchawie Lake (Zwanzig and Wielezynski, 1975) and at Russell Lake (Lenton, 1981). The new data suggest that, like the better preserved rocks discussed in the study by Zwanzig et al. (1999), the amphibolite comprises units with tectonic affinities to 1) arcs, 2) marginal-basin floors, and 3) ocean islands. They are considered to be, respectively, part of the 1) Laurie Lake assemblage, 2) Hatchet Lake and Tod Lake basalts, and 3) Pickerel Narrows amphibolite (as summarized in Table GS-17-1).

Wide variations of alkali elements and Ca within these units of amphibolite indicate that most of the rocks are altered to some extent. Consequently, plots of the least mobile elements are used to evaluate the data.

The assemblage types are distinguished on a simple plot of TiO_2 versus MgO that shows the fields of modern arc, MORB/BABB (back-arc basin basalt) and Hawaiian tholeiite (Fig. GS-17-3). The newly analyzed samples of Pickerel Narrows amphibolite from Granville Lake represent picrite of ocean-island affinity with a high content of TiO_2. High-Mg basalt that is depleted in TiO_2 is interpreted to be komatiitic (*see* 'Komatiitic Basalt and Felsite' section, below).

Laurie Lake Assemblage

Two new samples of low-Ti basalt were collected on the east shore of Laurie Lake (4010 on Fig. GS-17-2). They comprise highly flattened pillow basalt and porphyritic volcanic breccia, which is locally scoriaceous. The basalt lies stratigraphically beneath the Sickle Group basal conglomerate and above a coarse hornblende-bearing greywacke-turbidite. It occurs in a south-trending belt on the east side of the Laurie Lake dome, where lithologically similar rocks were recorded as units 4, 4b and 4f during the original mapping (Gilbert et al., 1980). The new samples suggest that the Laurie Lake arc assemblage extends for nearly 25 km along the east shore of this lake and further east. In an N-MORB–normalized extended-element plot, the low-Ti basalt has a similar profile to the main calc-alkaline part of the Laurie Lake assemblage (Fig. GS-17-4a); however, it contains less MgO and more Fe_2O_{3tot}, and has a higher Th/Nb ratio (Table GS-17-1). The basalt may represent an arc tholeiite belonging to the Fox mine succession, which overlies the calc-alkaline rocks at the extracted volcanogenic massive sulphide (VMS) deposit (Zwanzig, 1999).

Farther west on Laurie Lake, dark green amphibolite with prominent lenses and layers rich in coarse-grained diopside closely resembles highly recrystallized parts of the Laurie Lake assemblage. This rock is inverted on the north side of the Laurie Lake dome and may extend around the dome farther west toward Saskatchewan.

Hatchet Lake Basalt

Hatchet Lake basalt has a restricted compositional range and TiO_2/MgO ratios typical of N-MORB (Fig. GS-17-3), but is somewhat depleted in Zr (Table GS-17-1). It occurs in the southeast-facing structural sheet that overlies Burntwood Group turbidite southeast of the Laurie Lake assemblage (Zwanzig et al., 1999). Basalt from a highly flattened, dark grey weathering pillowed flow at a new sampling locality on Laurie Lake has a similar TiO_2/MgO ratio, as well as the characteristic, horizontal, MORB-normalized, extended-element profile (Fig. GS-17-4b). This rock lies in the upper part of the inverted structural package on the north side of the Laurie Lake dome and is cut off along the overturned unconformity at the base of the Sickle Group (4003 on Fig. GS-17-2).

Tod Lake Basalt

The most widespread map unit of basalt contains 1.0–2.5% TiO_2 and is classified as Tod Lake basalt (Fig. GS-17-3). It is Fe-rich and contains relatively high concentrations of Zr (Table GS-17-1). Zwanzig et al. (1999) suggested that it formed in a marginal basin by fractionation and contamination from a parent magma similar to that of the Hatchet Lake basalt. The contaminating crust or sedimentary rocks were probably heterogeneous and thus produced variable enrichment in light rare-earth elements (LREE) and depression of high field-strength element (HFSE)/rare-earth element (REE) ratios. Accordingly, the basalt shows trace-element variations between, and locally within, sampling areas. This is illustrated by La/Yb ratios (Table GS-17-1) and MORB-normalized extended-element plots (Fig. GS-17-4c to -4e). The new data from Granville Lake show a negative trace-element slope (Fig. GS-17-4d); data from widely scattered samples in South Bay, across the Laurie Lake dome, show a less prominent slope; and the old data (mainly from Tod Lake) show the most gentle slope (Fig. GS-17-4c). Rocks from a mafic sheet within the structural stack of Burntwood Group turbidite on Wheatcroft Lake lack a negative Nb anomaly (Fig. GS-17-4e).

Despite the variations in detail, the trace- and major-element compositions of Tod Lake basalt are similar over an area equivalent to that of the arc-related volcanic rocks in the Lynn Lake belt and probably extending east and west beyond the area of sampling.

Pickerel Narrows Amphibolite

Mafic to ultramafic rocks (Pickerel Narrows amphibolite of Zwanzig et al. [1999]) with exceptionally high contents of TiO_2 (Fig. GS-17-3), Zr and La/Yb ratios (Table GS-17-1) were obtained from sample collections that extend over 1500 m along the south shore of Granville Lake (Fig. GS-17-1). These rocks include alkali olivine basalt and picrite (?)flows with very consistent MORB-normalized extended-element profiles that have a steep negative slope (Fig. GS-17-4f). They are part of an ocean-island assemblage interlayered with the upper part of the Tod Lake basalt. The succession is overlain by a unit of ultramafic rock containing flattened fragments with variable textures and weathering colours. Rounded (?)boulders and angular clasts suggest that the unit may be a locally derived conglomerate. The relatively steep slope of the extended-element plots of some samples shows an affinity to the ocean-island assemblage.

Komatiitic Basalt and Felsite

High-Mg basalt with highly depleted contents of TiO_2, Zr and LREE (Fig. GS-17-3 and -4g) occurs in the structural stack on Granville Lake, and north of the Hatchet Lake basalt on Laurie Lake (4001 on Fig.

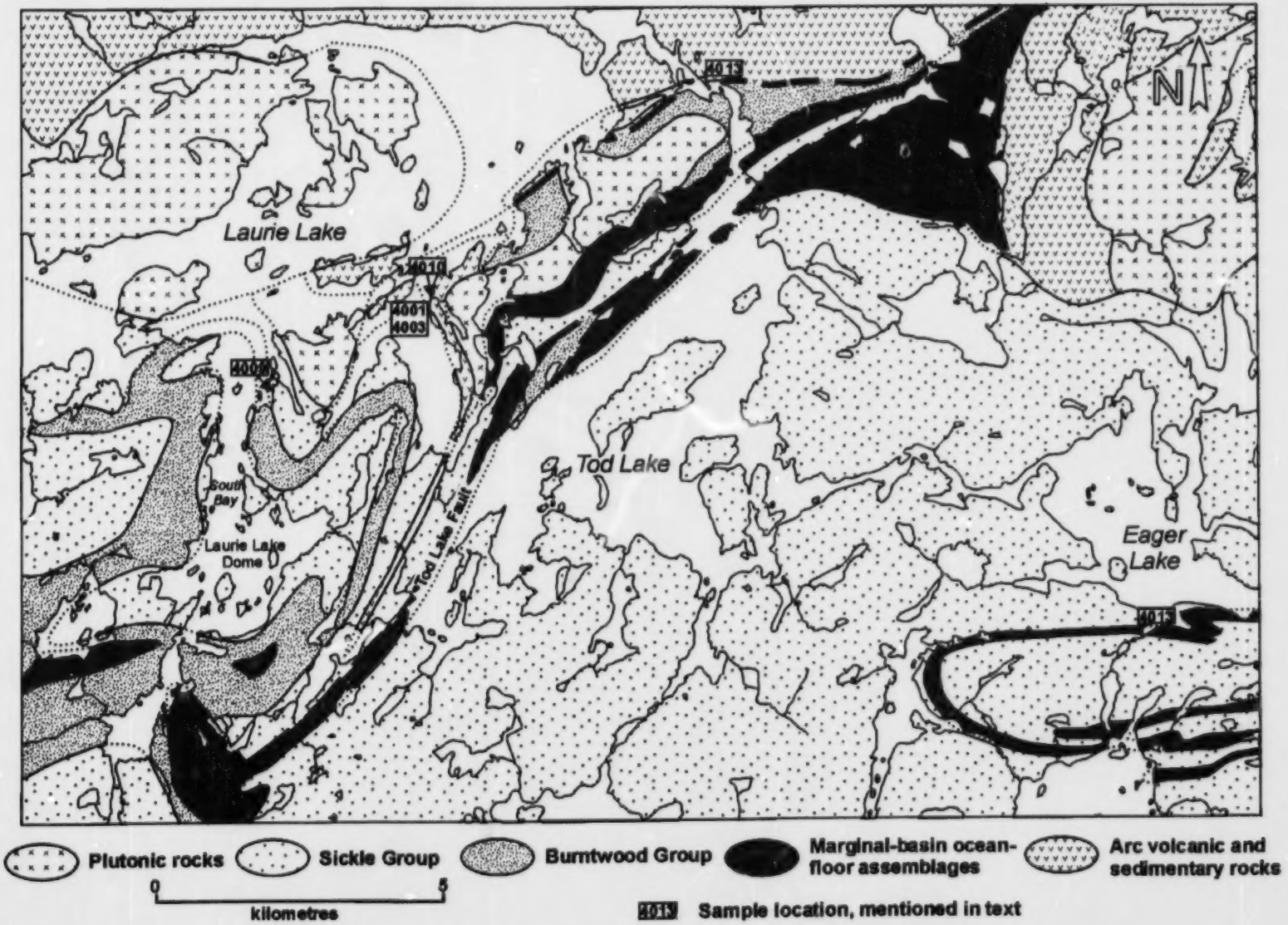

Figure GS-17-2: Geological map of the Laurie Lake–Eager Lake area (after Gilbert et al., 1980), showing the locations of selected geochemical samples.

Table GS-17-1: Typical and average analyses of selected elements and element ratios.

Sample	Unit	Source	SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O	TiO_2	TiO_2/MgO	La/Yb_N	Th/Nb_N	Zr
Average	Laurie Lake assemblage	OF-99-13	52.3	15.79	7.99	9.65	10.0	2.96	0.30	0.64	0.07	3.54	5.07	49
4010-1	Laurie Lake assemblage	Laurie L.	50.9	15.92	10.18	6.75	11.2	3.20	0.62	0.72	0.11	6.16	6.63	79
Average	Hatchet Lake basalt	OF-99-13	49.4	14.98	13.37	8.22	10.1	2.23	0.18	1.12	0.14	0.79	1.61	54
4003	Hatchet Lake basalt	Laurie L.	50.3	15.16	10.87	9.11	9.61	3.47	0.41	0.84	0.09	0.63		52
Average	Tod Lake basalt	OF-99-13	51.0	14.95	13.60	5.39	9.71	2.95	0.32	1.72	0.32	1.54	3.54	99
4009-1	Tod Lake basalt	South Bay	47.5	14.53	14.66	8.33	9.62	3.25	0.22	1.66	0.20	1.79	1.44	77
4015	Tod Lake basalt	Eager L.	48.3	14.03	14.03	6.50	11.5	2.67	0.63	1.91	0.29	2.80	4.37	118
606	Tod Lake basalt	Granville L.	51.4	15.69	11.56	5.65	10.2	2.83	0.60	1.72	0.30	3.18	5.68	113
772-8	Tod Lake basalt	Wheatcr L.	47.0	14.02	15.67	8.16	10.8	1.86	0.08	1.86	0.23	1.61	1.64	117
Average	Tod Lake type sills	Wheatcr L.	45.0	14.84	15.84	9.07	12.3	1.09	0.15	1.16	0.13	2.73	1.93	53
Average	Pickerel Narrows assembl	Granville L.	43.0	8.45	16.04	17.90	10.8	0.47	0.28	2.49	0.14	14.19	2.37	157
434-2	Pickerel Narr pillow basalt	Granville L.	42.3	14.69	16.29	8.82	10.9	2.42	0.60	3.37	0.38	6.91	1.62	192
Average	Granville Lake fragmental	Granville L.	48.8	8.52	11.15	18.58	10.5	0.94	0.09	1.07	0.06	3.96	3.83	63
4001	Komatiitic basalt	Laurie L.	47.3	15.47	12.93	9.82	11.7	1.74	0.30	0.46	0.05	0.29		18
553-2	Komatiitic basalt	Granville L.	48.6	12.88	9.17	15.17	11.8	1.65	0.44	0.10	0.01	0.39		6
4013-2	Pyroxenite	Laurie R.	48.5	6.68	13.18	16.19	13.6	0.86	0.23	0.33	0.02	3.55	7.83	25

Values in weight percent (volatile free), except for Zr, which is in ppm
$_N$–chondrite normalized

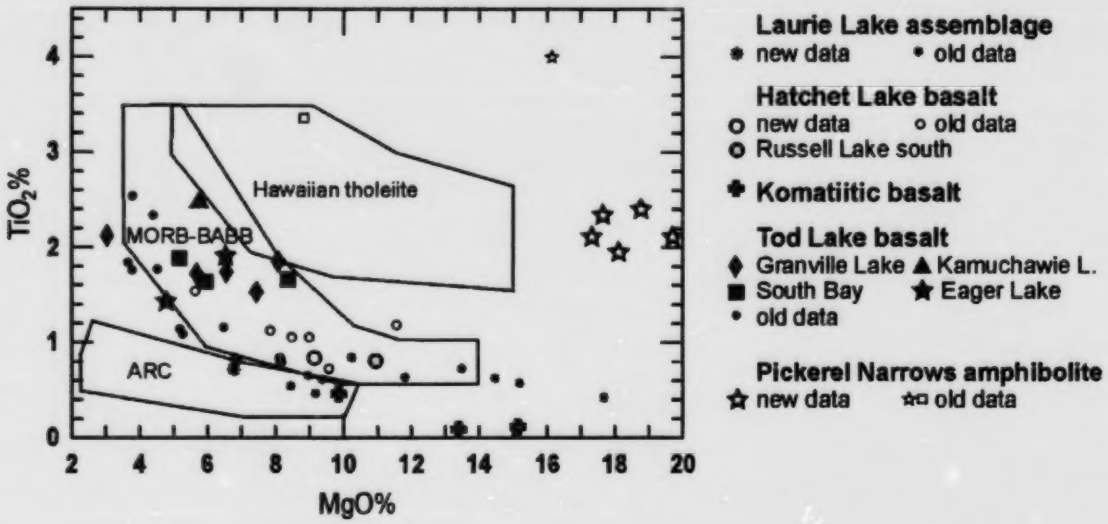

Figure GS-17-3: Plot of TiO_2 vs. MgO from mafic-ultramafic volcanic rocks in various tectonic assemblages along the Kisseynew Domain–Lynn Lake belt boundary (fields of modern tectonic environments after Stern et al. [1995]; old data from Zwanzig et al. [1999]).

GS-17-2). Its positively sloped MORB-normalized profile with negative Ti anomaly is typical of komatiitic basalt.

The komatiitic basalt occurrence on Laurie Lake is adjacent to the Laurie Lake felsite, which is a dyke or bed of Na-rhyolite tuff characterized by an intermediate to heavy REE profile with positive slope (Fig. GS-17-4g). The overall trace-element signature of the felsite has been related to arc rifting (Zwanzig et al., 1999). The Laurie Lake felsite may be part of a related bimodal suite, together with the komatiitic basalt.

Laurie River Ultramafic Intrusion

A small, highly recrystallized intrusion that probably represents pyroxenite and olivine pyroxenite occurs on the north shore of the Laurie River between Laurie Lake and Tod Lake. The poorly exposed body stratigraphically underlies the unconformity at the base of the Sickle Group, analogous to the Laurie Lake assemblage. Its negatively sloped, MORB-normalized, extended-element plot with negative Nb, Zr, Hf and Ti anomalies is very similar to the calc-alkaline Laurie Lake assemblage, which may be an extrusive equivalent of the Laurie River intrusion (Fig. GS-17-4h and -4a).

TECTONIC STACKING

Three of the lithotectonic assemblages of volcanic rocks occur in separate structural sheets, as indicated by younger sequences of the Sickle Group or Burntwood Group stacked between them. Contacts between assemblages are tectonic; only the Tod Lake basalt and the Pickerel Narrows amphibolite appear to be in stratigraphic contact, because thin (?)flow units are interlayered without evidence of faulting. Different tectonic environments are inferred for these assemblages based on the geochemical data (Zwanzig et al., 1999 and this report). They include volcanic-arc, marginal-basin, possible arc-rift and ocean-island settings. The assemblages, which are generally less than 300 m thick (in most areas <10 m thick), represent fault slivers of the original terranes. The structural pile that contains these slivers is the result of extensive telescoping and tectonic stacking; the faults between them must be major tectonic discontinuities.

A seismic profile across the east end of Granville Lake has been interpreted to show that the tectonic stack dips moderately northeast and extends through both the upper and lower crust (White et al., 2000). The hanging-wall block contains mainly the arc-related rocks of the Lynn

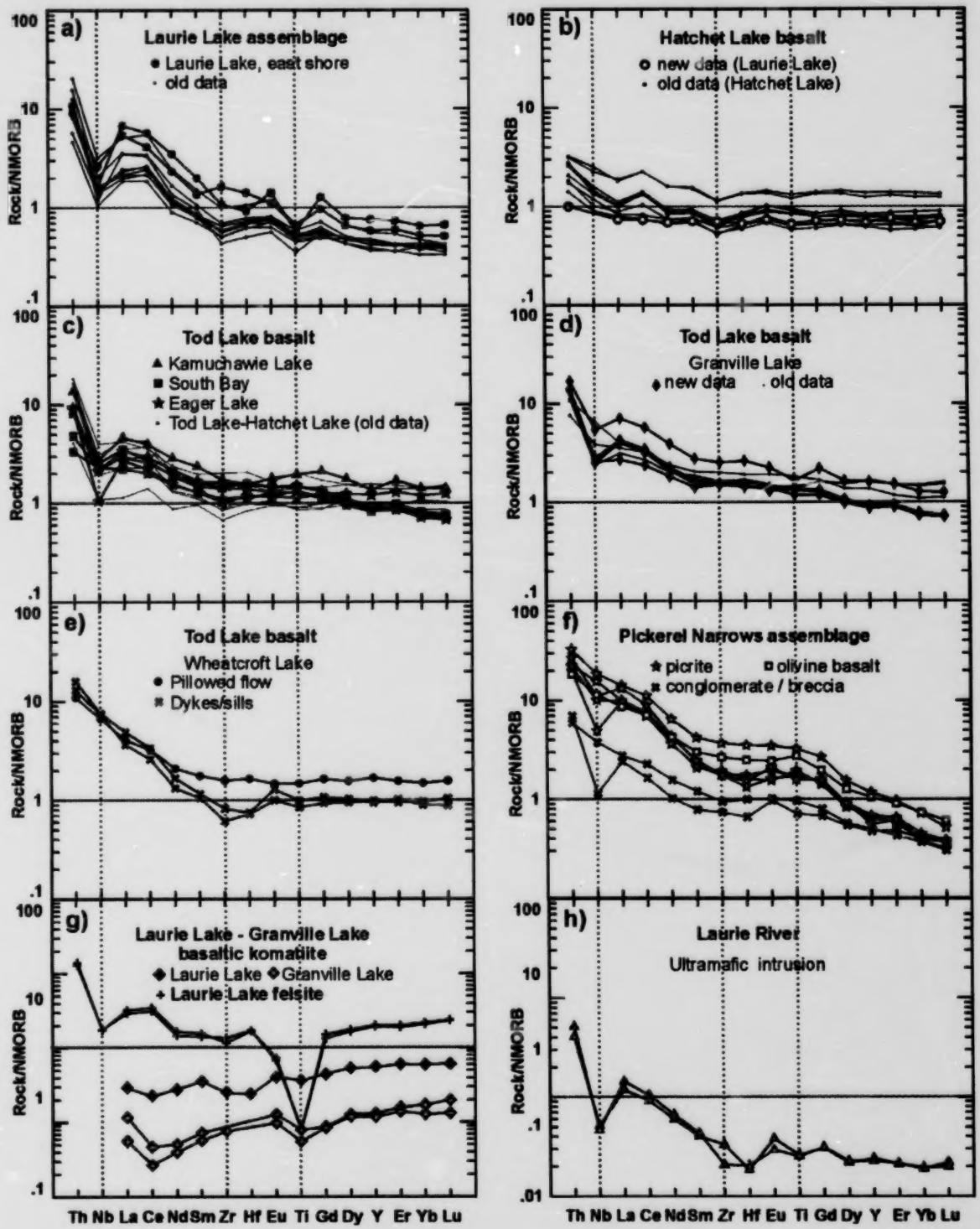

Figure GS-17-4: N-MORB–normalized extended-element plots from igneous rocks in various tectonic assemblages along the Kisseynew Domain–Lynn Lake belt boundary (normalizing values after Sun and McDonough [1989]; old data from Zwanzig et al. [1999]).

Lake belt, whereas the footwall is dominated by the younger Burntwood Group turbidite of the Kisseynew Domain. The crustal suture between these blocks is the Granville Lake Structural Zone (Zwanzig, 1990), which is dominated by the ocean-floor–marginal-basin assemblages that locally contain volcanic remnants of ocean-island affinity.

On Laurie Lake, the tectonic stack has been domed, highly flattened and refolded (Fig. GS-17-2). The structural relief due to the doming has provided a complete transect across the stack from South Bay, in the core of the Laurie Lake dome, northeast to the Laurie River, in the Lynn Lake belt. The deepest structural level, which is the main part of the dome, consists of large recumbent folds of the youngest rocks (the Burntwood and Sickle groups). Layered amphibolite, herein identified by geochemical criteria as Tod Lake basalt, separates the Burntwood and Sickle groups. Locally, hornblende-bearing quartzofeldspathic gneiss, derived from coarse-grained greywacke, lies between two fault-bounded sheets of Tod Lake basalt. Laminated quartzite, interpreted as chert, stratigraphically overlies the upper sheet of basalt and is, in turn, stratigraphically overlain by the Sickle Group. The whole structural package of amphibolite and gneiss is very similar to a section on Granville Lake, and represents the crustal suture exposed in the Granville Lake Structural Zone. At the deepest structural level, in South Bay, the zone is generally reduced by extreme attenuation to a few metres thickness. The deep part of the zone is easily recognized only on the west shore of South Bay, where a thickness of 200 m is preserved.

The sampled section of amphibolite on the north side of the dome contains a repetition of the suture zone at a higher structural level. This is the section in which the overturned unconformity at the base of the Sickle Group is underlain by Hatchet Lake basalt (MORB) and komatiitic basalt with Laurie Lake felsite (arc rift). Nearby, along strike, the unconformity is underlain by Laurie Lake assemblage (arc), along with hornblende-bearing greywacke. This section is up to 300 m thick. It is interpreted to contain three or four major tectonic breaks. An anticlinal core of Burntwood Group metasedimentary rocks to the northeast is bounded by another structural break. These attenuated terrane slivers and their tectonic discontinuities represent a structural style that is more complex than previously assumed. The style was produced by tectonic stacking, followed by ductile, polyphase deformation.

The timing of tectonic stacking and subsequent deformation are problems that require U-Pb zircon dating. The unconformity at the base of the Sickle Group overlies various arc and MORB-like units on the east shore of Laurie Lake and on adjacent islands at site 4001 (Fig. GS-17-2). The same relationship was reported at Tod Lake, and was interpreted to indicate that the accretion of the marginal-basin floor to the arc assemblages occurred early in the tectonic history, whereas subsequent thrusting involved the Sickle Group (White et al., 2000). Doming, refolding and final thrusting, such as along the Tod Lake Fault, occurred during peak and waning metamorphism. The late reactivation of shear zones produced zones of phyllonite and carbonate alteration that are well exposed on the west shore of Tod Lake. Displacement was probably connected with faulting along the Johnson Shear Zone in the central Lynn Lake belt.

APPLICATION TO MINERAL EXPLORATION

It has been shown in the Flin Flon Belt that the various tectonic assemblage types differ in mineral potential (Syme et al., 1999). An alteration zone with garnet, anthophyllite, gedrite and small amounts of Cu, Zn and Au (location 24 in Ferreira, 1993) occurs between Hatchet Lake basalt (MORB) and komatiitic basalt, along strike from the Laurie Lake felsite. More work is required to determine whether the Fox mine succession of arc tholeiite extends along the southeast shore of Laurie Lake, and possibly also occurs along the north shore at the Lar VMS deposit (Ferreira, 1993). The extreme attenuation of these units in the central and southern parts of Laurie Lake requires detailed mapping, in order to identify mineralized zones that are expected to be very thin. The surprisingly good preservation of trace- and rare-earth–element patterns makes lithogeochemistry an important tool for such mapping and for identifying favourable units.

The Caimito Au showing, on central Laurie Lake (location 25 in Ferreira, 1993), is hosted by deformed vein quartz in one of the high-strain zones that cut or reactivate the early tectonic breaks. Displacement along such zones may have been linked to that on the Johnson Shear Zone to the northeast. The zones lie in the wedge-shaped hanging-wall block above the Granville Lake Structural Zone, which probably served as a crustal-scale detachment zone (White et al., 2000) and as a possible channel for fluids feeding the shear zones.

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GS-18 U-PB AGE OF THE GABBRO AND OTHER PLUTONS AT LYNN LAKE (PART OF NTS 64C)

by A. Turek[1], J. Woodhead[2] and H.V. Zwanzig

Turek, A., Woodhead, J. and Zwanzig H.V. 2000: U-Pb age of the gabbro and other plutons at Lynn Lake (part of NTS 64C); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 97-104.

SUMMARY

The ore-hosting 'EL' and the barren Fraser Lake gabbroic stocks, which intrude the Wasekwan Group arc-related volcanic rocks in the Lynn Lake belt, have yielded U-Pb zircon ages of 1871.3 ± 2.4 Ma and 1870 ± 6.2 Ma, respectively. The Pb isotopic composition of the feldspars from the gabbro bodies indicates a common source with a μ value of 9.36. However, the barren Fraser Lake body has components derived from a different source with a higher but still mantle-like μ value. The Eden Lake granite, which intrudes gneissic tonalite at the southeast end of the Lynn Lake belt, has yielded an age of 1870 ± 10 Ma, comparable to the age of the gabbro bodies and to published ages for the Hughes Lake and Norrie Lake tonalite bodies. The age for a tonalite north of the Fox mine, at the southwest end of the belt, is significantly younger at 1831 ± 3.7 Ma. Thus, the period of intrusive activity is 1831 to 1871 Ma, similar to the span of 'successor arc' magmatism elsewhere in the Trans-Hudson Orogen.

INTRODUCTION

The Lynn Lake greenstone belt (Fig. GS-18-1) is a Paleoproterozoic segment of the Trans-Hudson Orogen (Hoffman, 1981). The arc-related volcanic rocks (Syme, 1985; Zwanzig et al., 1999) have yielded U-Pb zircon ages of 1915 +7/-6 and 1910 ± 10 Ma (Baldwin et al., 1987). These are intruded by the previously undated Lynn Lake gabbro bodies that host ore deposits (depleted) of Ni, Cu and Co (Pinsent, 1980). The gabbro bodes are intruded by the Pool Lake suite of granodiorite to diorite, which has yielded U-Pb ages of 1876 +8/-7 Ma for the Hughes Lake quartz diorite and 1876 +8/-6 for the Norrie Lake tonalite (Baldwin et al., 1987). The plutons and the Wasekwan Group were deformed and subsequently overlain by the Sickle Group fluvial-alluvial arenite (Table GS-18-1). Polyphase deformation and upper greenschist to upper amphibolite metamorphism have affected all the rocks in the Lynn Lake belt, including a suite of post-Sickle plutons.

A general lack of zircons in the rocks of the Lynn Lake belt has hampered this and other radiometric-dating programs. Nevertheless, we have been able to obtain precise ages for the early Lynn Lake gabbro, and to extend the dating to granite that was considered to be relatively early but yielded a surprisingly young age, showing that some of the tonalite bodies intruding the Wasekwan Group have a post-Sickle age. The U-Pb and Pb-Pb ages relevant to this report are listed in Appendix GS-18-1.

ANALYTICAL PROCEDURES

Samples for radiometric dating were approximately 30 kg, except for sample L12, which was approximately 80 kg. After crushing, zircons were separated by standard procedures using the Wilfley table, Carpco™ and Frantz™ magnetic separators, and ethylene tetrabromide and methylene iodide heavy liquids. Zircon fractions selected for analysis were upgraded by abrasion and hand picking, and cleaned thoroughly by ultrasonic vibration in 50% nitric acid and ultra-pure water. Sample dissolution followed the techniques of Krogh (1973) and Parrish (1987), and samples were spiked with a mixed ^{208}Pb and ^{235}U tracer for the concentration measurements. The feldspar samples were separated similarly, using magnetic separators and heavy liquids. They were dissolved in Savillex™ Teflon™ vessels and spiked with natural ^{203}Tl-^{205}Tl to act as an internal standard.

The Pb and U isotopic ratios were measured on a 90°, 25 cm mass spectrometer. The composition of Pb was measured on unspiked aliquots, and U and Pb concentrations and the U/Pb isotopic ratios were calculated from spiked aliquot runs; uncertainties in the U/Pb ratios are ±1.0% at the 95% confidence level. Isotopic ratios were corrected for mass fractionation, blank and initial common Pb. The mass-fractionation correction used was 0.10% per atomic mass unit. The Pb-blank correction was 12 or 70 pg Pb, depending on the size of the Teflon™ pressure bomb used. The composition of the blank used was $^{206}Pb/^{204}Pb$ = 18.7, $^{207}Pb/^{204}Pb$ = 15.7 and $^{208}Pb/^{204}Pb$ = 38.4. Correction for the initial common Pb present was done using the Stacey and Kramers (1975) equations. The Pb-isotope composition of the feldspar samples was measured on an Nu Instruments™ 'Nu plasma' multicollector ICP–MS instrument. Mass fractionation of the Pb isotopes was corrected on the basis of mass bias observed for the added Tl internal standard of known isotopic composition.

Data reduction for the zircon samples followed the procedure of Ludwig (1982). Statistical evaluation of the results was done using the plotting and regression analysis of Ludwig (1999). For the U-Pb data, the regression provides the upper and lower concordia age intercepts and also provides two measures of goodness of fit: the mean square of weighted deviates (MSWD), which is a residual variance, and a probability of fit. The ages being reported here are within experimental error and the uncertainties are at the 95% confidence limit.

RESULTS

Of the 16 rock units sampled, only four contained zircons. The locations of all samples are shown on Figure GS-18-1 and their co-ordinates are given in Appendix GS-18-2. The analytical data are given in Tables GS-18-2 and -3, and are shown as concordia diagrams (Fig. GS-18-2 to -5) and as Pb-evolution plots (Fig. GS-18-6).

'EL' Gabbro (L12)

The 'EL' gabbro intrudes mafic to intermediate porphyritic flows of the high-LREE (light rare-earth elements) arc tholeiite (Wasekwan Group) in the northern belt (Zwanzig et al., 1999). This intrusion is a vertical pipe, 500 m in diameter, and hosts a high-grade Ni-Cu deposit that has been mined out from the 700 m level up to the surface. This stock is probably connected to the larger 'A' stock at the Lynn Lake townsite, which hosts several Ni-Cu orebodies. A sample of this gabbro yielded clear, colourless to light yellow zircons that were euhedral with some overgrowths, which were removed by abrasion. Five fractions were analyzed: one is concordant; three are discordant by less than 2%; and one plots above the concordia, which we attribute to the abrasion process. All five data sets are collinear within experimental limits and define a concordia age of 1871.3 ± 2.4 Ma with a lower intercept of 291 ± 183 Ma.

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Figure GS-18-1: Geological sketch map showing location of samples (modified from *Beaumont-Smith and Rogge, 1999*).

Table GS-18-1: Table of formations (adapted from Gilbert et al., 1980). Unit numbers are those used by Gilbert et al. (1980).

Supracrustal rocks		Intrusive rocks			U-Pb age	
Unit	Rock type	Unit	Rock type	Unit	Rock type	Age (Ma)
		Post-Sickle intrusive rocks				
		18-22	Diorite, tonalite, granite, granodiorite, pegmatite, porphyry			
		16	Tonalite, quartz diorite	16	Fox Mine tonalite[b]	1831 ± 3.7
Sickle Group						
11-12	Conglomerate, sandstone and gneiss					
Burntwood Group						
1	Greywacke, mudstone, siltstone					
	Sickle or Wasekwan Group					
10	Conglomerate with sedimentary, volcanic and granitoid clasts					
Unconformity						
		Pre-Sickle intrusive rocks				
		16-17	Diorite, tonalite, granite, granodiorite, aplite, pegmatite	17 16 16	Eden Lake granite[b] Hughes Lake diorite[a] Norrie Lake tonalite[a]	1871 ± 10 1876 ± 7.5 1876 ± 7
		13-15	Gabbro,diorite, norite, diabase	15 15	Fraser Lake gabbro[b] 'EL' gabbro[b]	1870 ± 6.2 1871 ± 2.4
Wasekwan Group						
8-9	Conglomerate, coarse- to fine-grained sedimentary rocks, paragneiss					
5-7	Intermediate and felsic volcanic rocks, dacite, rhyolite and felsic gneiss			7	Lynn Lake rhyolite[a]	1915 ± 6.5 1910 ± 12
2-4	Aphyric basalt, porphyritic basalt, mafic and intermediate volcanic rocks, amphibolite					

[a] Age from Baldwin et al. (1987).

[b] Age from this study.

Table GS-18-2: Analytical data for zircons from Lynn Lake.

Sample detail[a]				Concentration		Atomic ratios					Apparent ages (Ma)[e]		
				(ppm)		(b)	(c)	(c)	(d)	(d)			
Sample No.	Magnetism	Grain size(μm)	Weight (mg)	U	Pb	$^{204}Pb/^{206}Pb$	$^{208}Pb/^{206}Pb$	$^{207}Pb/^{206}Pb$	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$	$^{208}Pb/^{232}Th$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
'EL' gabbro (L12)													
A	m0	114	0.3	76	42	0.01001	0.46768	0.22960	5.156	0.3276	1827	1845	1866
B	m0	114	0.4	75	41	0.00785	0.43719	0.20498	5.461	0.3456	1914	1895	1873
C	m0	114	0.2	122	46	0.00186	0.18673	0.11518	5.338	0.3377	1876	1875	1874
D	m0	62	1.9	251	90	0.00011	0.18348	0.11410	5.066	0.3223	1801	1830	1864
E	m0	114	1.1	225	82	0.00019	0.17930	0.11411	5.146	0.3270	1824	1844	1866
Fraser Lake gabbro (L11)													
A	m0	114	2.0	286	107	0.00085	0.18923	0.12450	5.161	0.3281	1829	1846	1865
B1	m4	114	1.3	472	169	0.00159	0.23638	0.13363	4.636	0.2969	1676	1756	1852
CC	m0	62	8.0	350	107	0.00048	0.15509	0.11836	4.289	0.2776	1579	1691	1833
CX	m0	62	2.7	451	156	0.00024	0.13791	0.11362	4.966	0.3170	1775	1813	1858
Eden Lake granite (L10)													
A	m1	114	2.3	618	215	0.00060	0.12768	0.12096	5.012	0.3209	1794	1821	1853
B	m2	114	1.1	550	169	0.00053	0.12385	0.11616	4.333	0.2853	1618	1700	1802
C	m2	114	0.9	778	246	0.00034	0.11601	0.11489	4.559	0.2971	1677	1742	1821
D	m1	114	2.0	746	232	0.00024	0.10368	0.11377	4.528	0.2959	1671	1736	1816
Fox mine tonalite (L5)													
B	m0	62	1.4	367	135	0.00174	0.12507	0.13427	5.109	0.3312	1844	1838	1830
C	m0	114	0.3	427	142	0.00108	0.09824	0.12077	4.809	0.3135	1758	1787	1820
D	m0	62	3.6	494	163	0.00038	0.07817	0.11643	4.915	0.3193	1786	1805	1827
E	m1	62	1.1	858	244	0.00036	0.07228	0.11300	4.167	0.2775	1579	1668	1781

[a] Relative magnetic susceptibility of zircons is reported as m0 (nonmagnetic) to m4 (paramagnetic) and is related to the inclination of the Frantz isodynamic separator using maximum current of 2 A. Grain size is an average; sieves used were 85, 150, 75, 38 μm.

[b] Measured ratio.

[c] Blank corrected.

[d] Blank and nonradiogenic Pb corrected.

[e] Decay constants used: $\lambda^{238}U = 1.55125 \times 10^{-10}$ year^{-1}; $\lambda^{235}U = 9.8485 \times 10^{-10}$ year^{-1} (Steiger and Jäger, 1977).

Table GS-18-3: Isotopic composition of Pb in feldspars from gabbroic stocks at Lynn Lake.

Sample detail	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
'A' stock gabbro (L2)	15.639	15.201	35.134
'A' stock gabbro (L7)	15.663	15.215	35.516
Fraser Lake tonalite (L4)	15.905	15.319	35.426
Fraser Lake gabbro (L11)	15.994	15.321	35.380
Fraser Lake gabbro (L8)	15.844	15.258	35.308
'EL' stock gabbro (L3)	16.110	15.281	35.477
'EL' stock gabbro (L12)	17.205	15.415	35.708

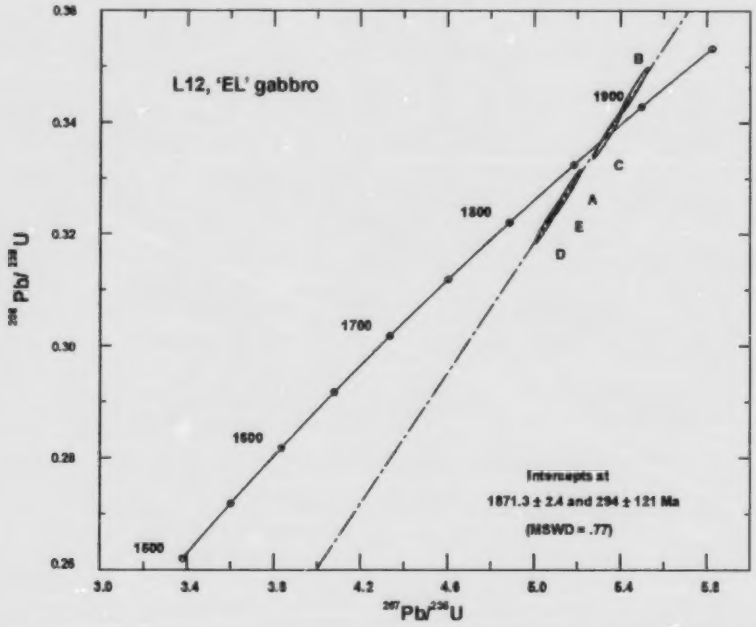

Figure GS-18-2: Concordia diagram for the 'EL' gabbro. A, B, C, D and E are separate zircon fractions.

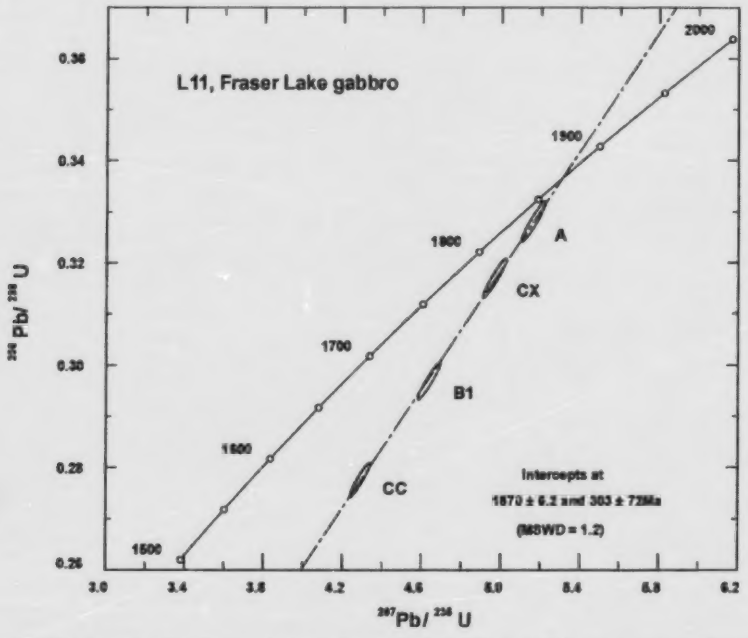

Figure GS-18-3: Concordia diagram for the Fraser Lake gabbro. A, B1, CC and CX are separate zircon fractions.

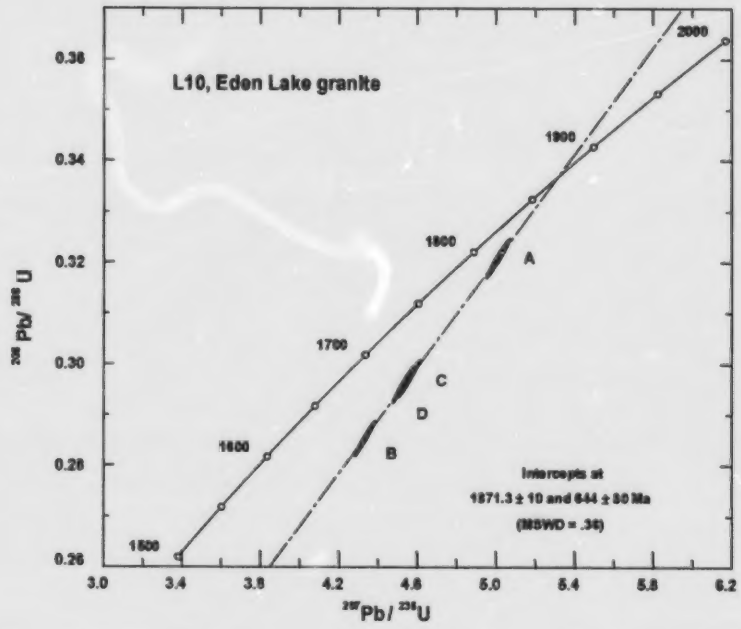

Figure GS-18-4: Concordia diagram for the Eden Lake granite. A, B, C and D are separate zircon fractions.

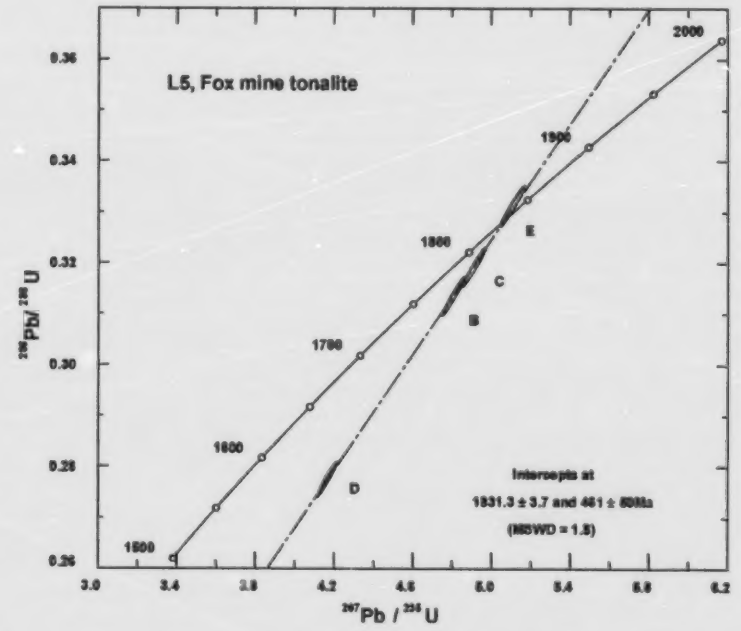

Figure GS-18-5: Concordia diagram for the Fox mine tonalite. B, C, D and E are separate zircon fractions.

Fraser Lake Gabbro (L11)

This pluton is a gabbro similar to the 'EL' and 'A' bodies, but it does not contain any sulphide mineralization. It intrudes the core of an early anticline and stitches rocks in the northern belt to those in the southern belt. The zircons separated from the gabbro sample are transparent and colourless to light pink. The four fractions analyzed are 2 to 20% discordant, but are collinear within experimental error and define a concordia age of 1870 ± 6.2 Ma with a lower intercept of 303 ± 72 Ma. The minimum $^{207}Pb/^{206}Pb$ age is 1865 Ma.

Eden Lake Granite (L10)

A sample of K-feldspar porphyritic granite from a batholith to the southeast of the main greenstone belt yielded a heterogeneous population of euhedral zircons. There appear to be two populations of zircons, cloudy white opaque crystals and orange-brown and iron-stained opaque crystals. Four fractions were analyzed: A is a strongly abraded and D a slightly abraded mixture of both populations, whereas B and C are not abraded but are hand-picked fractions of the white and orange-brown crystals, respectively. As is evident from the concordia plot, the data points are collinear but discordant by 6 to 22%. The concordia age is

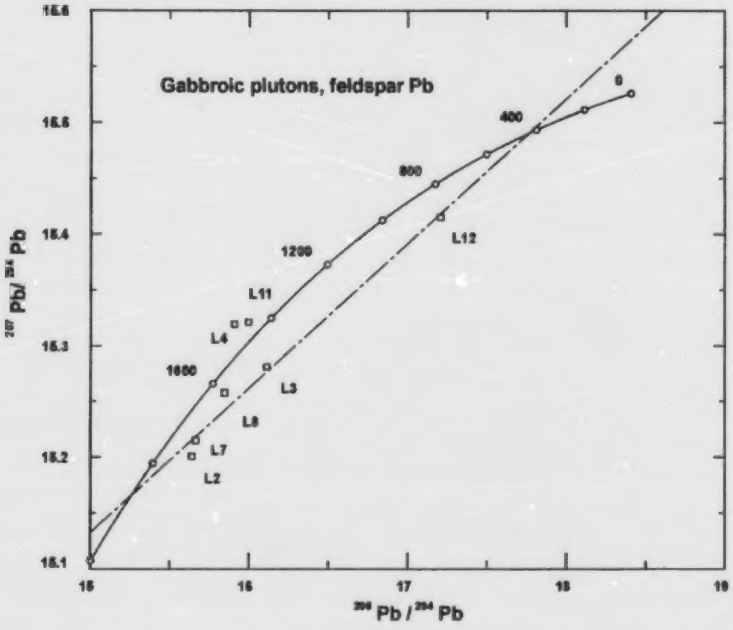

Figure GS-18-6: Pb isotopic composition of feldspars from the gabbro plutons. The Pb-evolution curve has a μ value of 9.36. Growth curve intercepts are 1870 and 430 Ma. The MSWD of the regression is 0.957. Samples L2 and L7 are from the 'A' stock; samples L3 and L12 are from the 'EL' stock; and samples L8, L4 and L11 are from the Fraser Lake stock.

1870 ± 10 Ma with a lower intercept of 644 ± 80 Ma. The minimum $^{207}Pb/^{206}Pb$ age is 1853 Ma.

Fox Mine Tonalite (L5)

This body, granodiorite to tonalite in composition, cuts off the Hatchet Lake mid-ocean ridge (MORB)-like basalt (Zwanzig et al., 1999), a narrow unit of sedimentary rocks, and the fault between them. Zircons separated from a tonalite sample are stubby, transparent and pink. Fractions B and C are strongly abraded, and fractions D and E are lightly abraded. Fraction E is concordant, whereas B, C and D are 3 to 20% discordant. The data define a discordia within experimental error. The indicated age is 1831 ± 3.7 Ma with a lower intercept of 461 ± 50 Ma.

Gabbro Plutons, Feldspar Analysis

The isotopic composition of Pb was measured on feldspars separated from the following samples: L2, L3, L4, L7, L8, L11 and L12. Feldspars were used because they concentrate Pb and therefore retain low U/Pb ratios. ICP–MS analysis of the feldspars used here indicates the presence of some U. Nevertheless, these analyses represent the best chance of approaching initial Pb isotopic compositions. The isotopic compositions are given in Table GS-18-3 and shown as a Pb-evolution diagram in Figure GS-18-6.

DISCUSSION

The four U-Pb zircon ages measured here, together with the four ages published by Baldwin et al. (1987), provide a useful though far from adequate control on the evolution of the Lynn Lake greenstone belt. The 'A' and 'EL' gabbro stocks carry economic sulphide mineralization, while the Fraser Lake gabbro, which has been extensively explored by geophysical methods and drilling by Sherritt Gordon Mines Ltd., is barren. The ages obtained here for the 'EL' and Fraser Lake gabbro bodies are 1871 ± 2.4 and 1870 ± 6.2 Ma, respectively. These ages are statistically identical at the 95% level and are also indistinguishable from those for the Pool Lake suite tonalite, which intrudes the gabbro. This important intrusive event must be approximately 5 Ma younger than the felsic volcanism in the adjacent Rusty Lake greenstone belt, with which it was previously thought to be coeval (Baldwin et al., 1987). The 1870 ± 10 Ma age of the Eden Lake granite, reported here, is also statistically part of this intrusive activity. Cameron (1988) has shown that the Eden Lake granite is relatively late in the sequence of igneous activity. Its pre-Sickle age suggests that the main pulse of arc magmatism may have waned earlier in the Lynn Lake–Rusty Lake region than previously thought. However, the age obtained for the Fox mine tonalite is much younger at 1831 ± 3.7 Ma. This age is the same as the 1832 Ma age for late-tectonic quartz diorite at Veronica Lake, to the north of the area (Van Schmus and Schledewitz, 1986).

The Pb isotopic composition of feldspars separated from the gabbroic stocks shows that the ore-hosting 'A' and 'EL' stocks both lie on a straight line that intersects the growth curve with a μ value of 9.36 at 1870 Ma and are therefore probably derived from the same source. Feldspars from the Fraser Lake body are from gabbro (L11 and L8) and tonalite (L4). Sample L8 lies, within error, on the same chord, but the other two Fraser Lake samples are above that curve. This suggests that the Fraser Lake body may be derived from two sources, one being the same as that of the 'A' and 'EL' stocks (with a μ value of 9.36) and another having a higher but still mantle-like μ value. Thus, there is some difference isotopically between the ore-hosting gabbroic stocks and the barren Fraser Lake stock.

The age determinations reported here are consistent with those of Baldwin et al. (1987) but suggest that the main 'successor-arc' magmatism in the Lynn Lake region may have peaked at ca. 1.87 Ga, earlier than elsewhere in the Trans-Hudson Orogen. The isotopic composition of the feldspars indicates a mantle origin, with no crustal contribution. This also supports the paleomagnetic evidence of Dunsmore and

Symons (1990) and Symons (1991), which indicates a wide separation of the Superior and Hearne–Rae cratons during the main stage of formation of the Lynn Lake greenstone belt.

The Fox mine pluton provides an important minimum age for the fault between the Hatchet Lake basalt and the Burntwood Group turbiditic metasedimentary rocks to the northwest (Zwanzig et al., 1999). The fault has been interpreted as part of a system of crustal-scale structures along which the turbidites in the Kisseynew Domain underthrust the arc volcanic rocks in the Lynn Lake belt (White et al., 2000). With an age of greater than 1831 ± 3.7 Ma, the faulting is coeval with the thrusting between the Burntwood Group and the volcanic rocks in the Flin Flon greenstone belt (Connors, 1996).

CONCLUSIONS

A limited number of U-Pb zircon ages suggest that the bulk of plutonic activity, mafic and felsic, in the Lynn Lake area took place around 1870 to 1876 Ma. The emplacement of the 'A' and 'EL' gabbro plutons with their massive sulphide orebodies took place at 1871 ± 2.4 Ma. The larger but barren Fraser Lake gabbro has a similar age of 1870 ± 6.2 Ma. While all three gabbro bodies are derived from a common mantle source with a μ value of 9.36, the Fraser Lake pluton also contains material derived from a mantle source with a higher μ value. These ages are not significantly different from the 1876 Ma ages reported for the Norrie Lake tonalite and the Hughes Lake tonalite (Baldwin et al., 1987), as well the 1871 ± 10 Ma age of the Eden Lake granite. The younger 1831 ± 3.7 Ma age of the Fox Lake tonalite constrains the timing of faulting at the tectonically active south margin of the Lynn Lake belt.

ACKNOWLEDGMENTS

We thank R. Reid and R. Deklerk for help with sample collection, I. Churchill for assistance with the chemistry, and J. Jurkowksi for the feldspar mineral separation. This study also acknowledges financial assistance from the LITHOPROBE Project and other Natural Sciences and Engineering Research Council (NSERC) grants.

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Appendix GS-18-1: Compilation of U-Pb and Pb-Pb isotopic ages for the Lynn Lake area: Lynn Lake Domain, South Indian Domain, Leaf Rapids Domain.

Lynn Lake Domain			
Eden Lake granite	U-Pb	1871 ± 10	this study
Fox mine tonalite	U-Pb	1831 ± 3.7	
Fraser Lake gabbro	U-Pb	1876 ± 7.5	
'EL' gabbro	U-Pb	1871 ± 3.5	
Rhyolite (FL1)	U-Pb	1915 +7/-6	Baldwin et al. (1987)
Rhyolite (FL2)	U-Pb	1910 +15/-10	
Pre-Sickle intrusive rocks			
Hughes Lake quartz diorite (3002)	U-Pb	1876 +8/-7	Baldwin et al. (1987)
Norrie Lake tonalite (3005)	U-Pb	1876 +8/-6	
South Indian Domain			
Veronica Lake			
Quartz diorite (MAN84-4)	U-Pb	1832	Van Schmus and Schledewitz (1986)
Leaf Rapids Domain			
Rusty Lake Belt			
Karsakuwingmak rhyolite (R1)	U-Pb	1878 ± 3	Baldwin et al. (1987)
Karsakuwingmak rhyolite (R2)	U-Pb	1874 +8/-7	
Ruttan mine, Leaf Rapids			
Sulphide ore (SP-2318)	Pb-Pb	1768	Sangster (1978)

Appendix GS-18-2: Sample location co-ordinates.

Sample	Unit	Latitude	Longitude
L2	Lynn Lake 'A' plug, hornblende-quartz gabbro	56°51'09"N	101°02'12"W
L3	Lynn Lake 'EL' plug, hornblende gabbro	56°49'14"N	101°01'55"W
L4	Fraser Lake tonalite	56°48'23"N	101°06'34"W
L5	Fox mine tonalite	56°38'25"N	101°38'41"W
L7	Lynn Lake 'A' plug, hornblende gabbro	56°51'05"N	101°02'00"W
L8	Fraser Lake hornblende-quartz gabbro	56°47'49"N	101°05'12"W
L10	Eden Lake porphyritic granite	56°41'15"N	99°57'06"W
L11	Fraser Lake quartz-hornblende gabbro	56°47'49"N	101°05'12"W
L12	Lynn Lake 'EL' plug, gabbro	56°49'14"N	101°01'55"W

GS-19 OPERATION SUPERIOR: MULTIMEDIA GEOCHEMICAL SURVEYS IN THE KNEE LAKE GREENSTONE BELT (NORTHERN HALF), NORTHERN SUPERIOR PROVINCE, MANITOBA (NTS 53L)

by M.A.F. Fedikow and E. Nielsen

Fedikow, M.A.F. and Nielsen, E. 2000: Operation Superior: multimedia geochemical surveys in the Knee Lake greenstone belt (northern half), northern Superior Province, Manitoba (NTS 53L); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 105-107.

SUMMARY

During the field season, 2000 rock, till, b-horizon, humus and vegetation samples were collected for multi-element geochemical analysis from 350 sites within the northern portion of the Knee Lake greenstone belt. In addition, 334 bulk samples of till and beach sand were collected for kimberlite indicator-mineral identification and analysis. The geochemical and mineralogical analysis of these samples will continue to build a multi-element–multimedia geochemical database that was initiated in 1996 and continued in 1997, 1998 and 1999 (Fedikow et al., 1996, 1997a, b; Fedikow and Nielsen, 1997; Fedikow et al., 1998, 1999, 2000) to assist mineral resource assessment in the northern Superior Province of Manitoba.

1999 SURVEY

The fifth and final year of a helicopter- and fixed-wing–assisted, belt-scale (1 km sample spacing), multimedia geochemical survey in northeastern Manitoba concentrated on the northern portion of the Knee Lake greenstone belt (Fig. GS-19-1). The distribution of sampling sites for the 2000 survey is given in Figure GS-19-2. Sample collection throughout the area was accomplished by helicopter, and a boat was used to access the extensive shoreline at the north end of Knee Lake. Samples are currently undergoing preparation for geochemical analysis in the laboratories of the Manitoba Geological Survey.

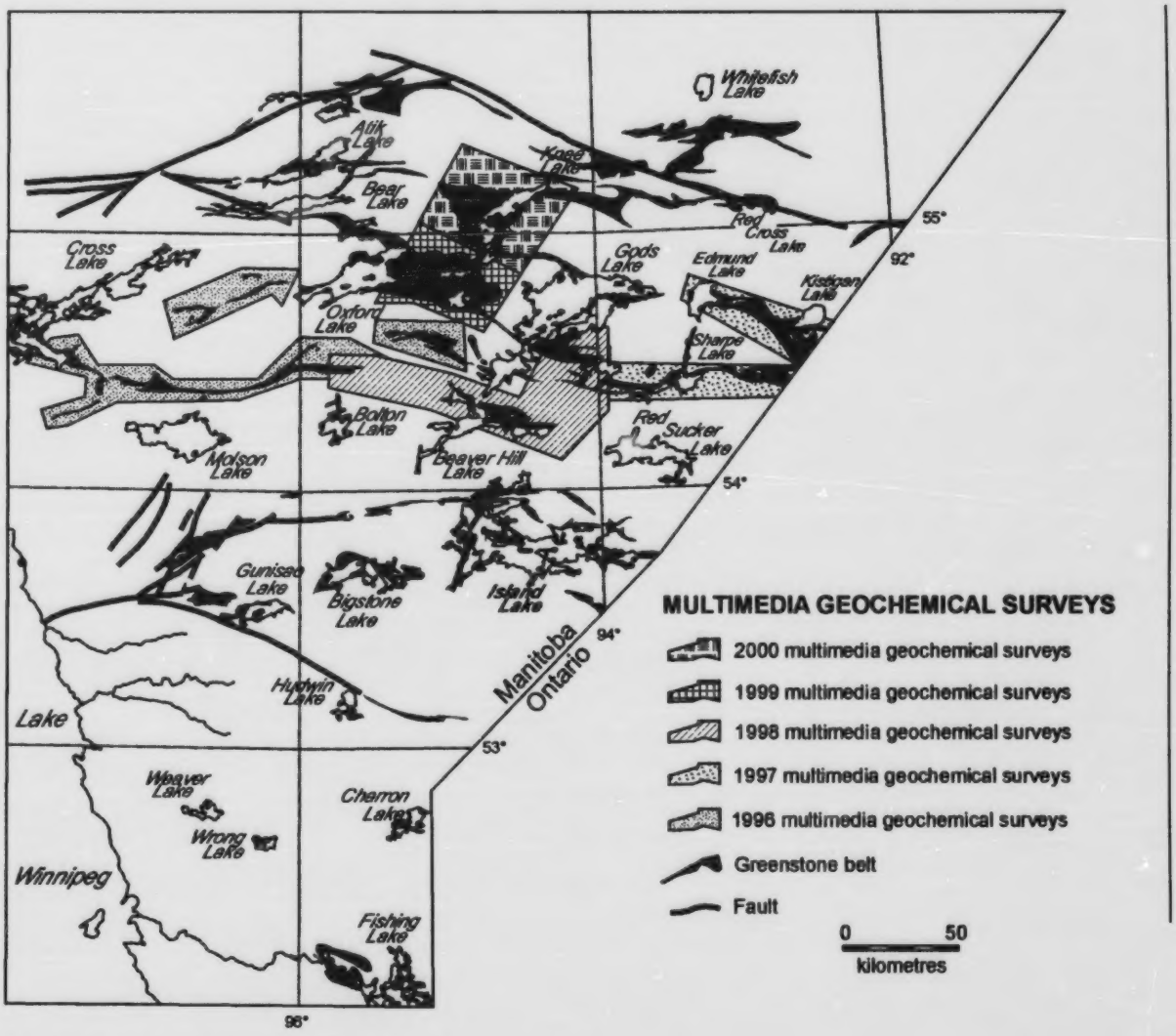

Figure GS-19-1: Target area for multimedia geochemical surveys, 2000.

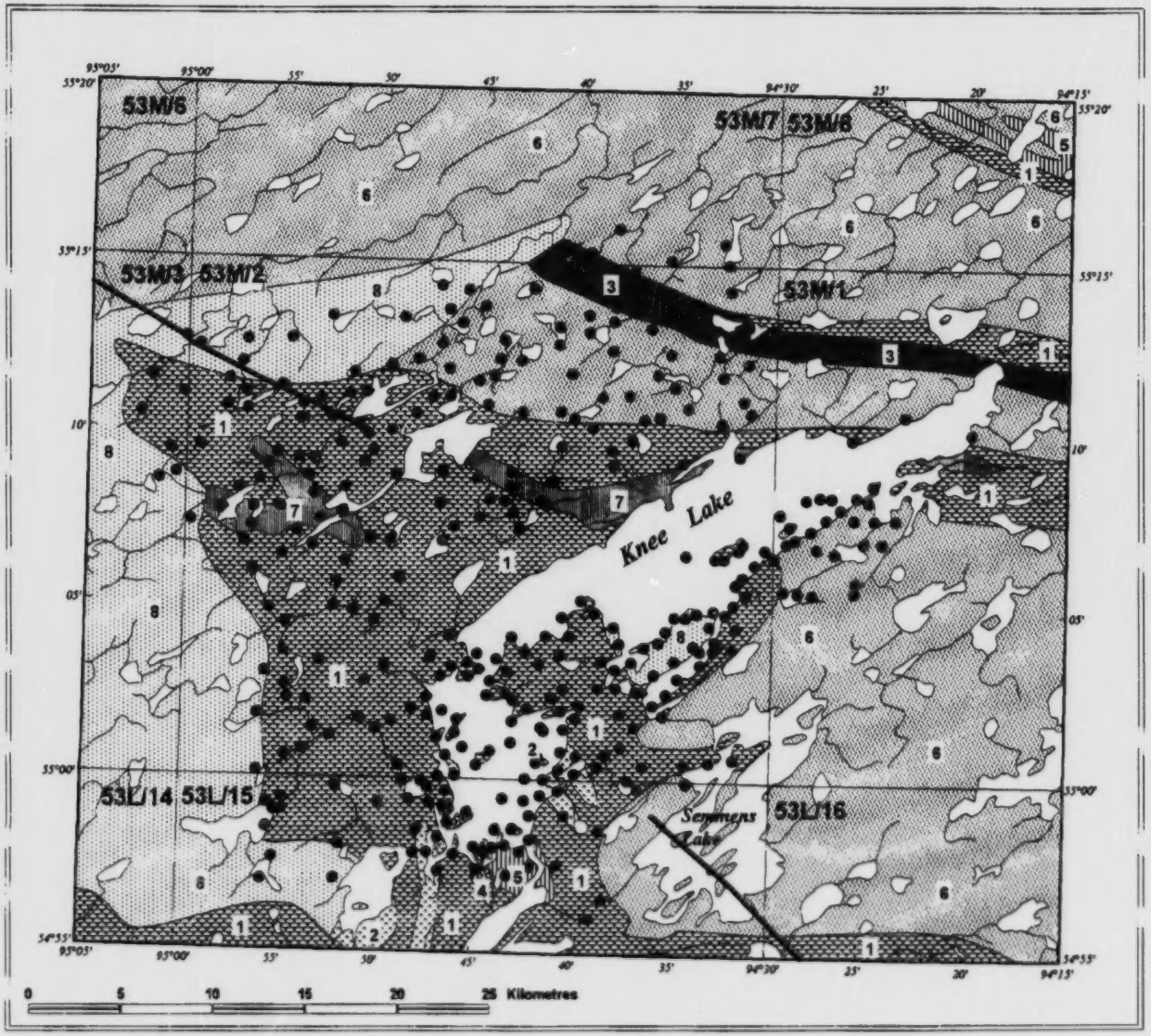

Figure GS-19-2: Distribution of sampling sites for the 2000 survey area.

ACKNOWLEDGMENTS

We gratefully acknowledge the considerable skills of Dave Tomlin and Jamie Gould, Provincial Helicopters Ltd. (Lac du Bonnet) in safely accessing sample sites for the 1999 season. Chris Bater, Zoe Kepron and Ryan Oger provided able and enthusiastic assistance during the sampling program and sample preparation. Neill Brandson and Karen Punstel are thanked for all-around logistical support.

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GS-20

WESTERN SUPERIOR NATMAP: AN INTEGRATED VIEW OF ARCHEAN CRUSTAL EVOLUTION

by J.A. Percival[1], A.H. Bailes, M.T. Corkery, B. Dubé[2], J.R. Harris[1], V. McNicoll[1], D. Panagapko[1], J.R. Parker[3], N. Rogers[1], M. Sanborn-Barrie[1], T. Skulski[1], D. Stone[3], G.M. Stott[3], K.Y. Tomlinson[1], J.B. Whalen[1] and M.D. Young[4]

Percival, J.A., Bailes, A.H., Corkery, M.T., Dubé, B., Harris, J.R., McNicoll, V., Panagapko, D., Parker, J.R., Rogers, N., Sanborn-Barrie, M., Skulski, T., Stone, D., Stott, G.M., Tomlinson, K.Y., Whalen, J.B. and Young, M.D. 2000: Western Superior NATMAP: an integrated view of Archean crustal evolution; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 108-116.

INTRODUCTION

The western Superior NATMAP project aims to provide a modern geoscience synthesis of northwestern Ontario and eastern Manitoba using resources of the federal, Ontario and Manitoba Geological Surveys, working at a variety of detailed and regional map scales. Key areas have been selected to address questions regarding relationships between Mesoarchean and Neoarchean sequences of the western Superior Province (Fig. GS-20-1; Percival et al., 1997), through new mapping, geochemistry, geochronology and GIS-based compilation. In 2000, the fourth year of field operations, focus was on the Uchi Subprovince (Fig. GS-20-2) at the southern margin of the Mesoarchean North Caribou Terrane, with continued work on the Wabigoon and Sachigo subprovinces.

In this paper, we present highlights of 2000 field and ongoing laboratory studies. Further syntheses in the form of a series of compilation maps will be released over the next year.

UCHI SUBPROVINCE

Lake Winnipeg Area

A collaborative mapping program in the Black Island area (Fig. GS-20-2) followed initial reconnaissance studies (Bailes, 1999; Percival and Whalen, 2000). At least three distinct supracrustal sequences are present, in addition to a ca. 3 Ga basement complex.

Basement complex

Hornblende-biotite tonalite of the East Shore complex represents basement to the Lewis–Storey assemblage. These coarse-grained, homogeneous rocks with characteristic blue quartz resemble tonalitic units in the vicinity dated at 3006 to 2998 Ma (Ermanovics, 1970, 1981; Krogh et al., 1974; Turek and Weber, 1994; V. McNicoll, unpublished data, 2000). North-trending, chlorite-grade shear zones (D_1) carry distinctive shear-band fabrics indicating dextral strike-slip motion.

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[3] Ontario Geological Survey, 933 Ramsey Lake Road, Sudbury, ON P3E 6B5
[4] Department of Geological Sciences, Queen's University, Kingston, ON K2B 6N1

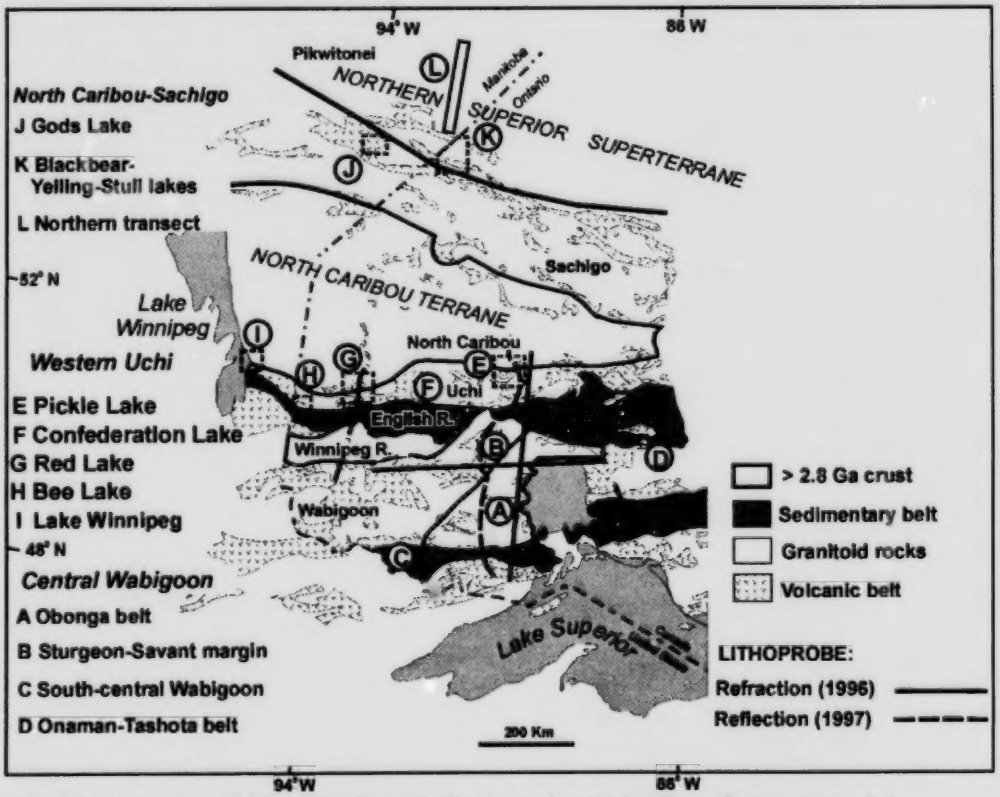

Figure GS-20-1: Schematic tectonic map of the western Superior Province, showing locations of areas under study.

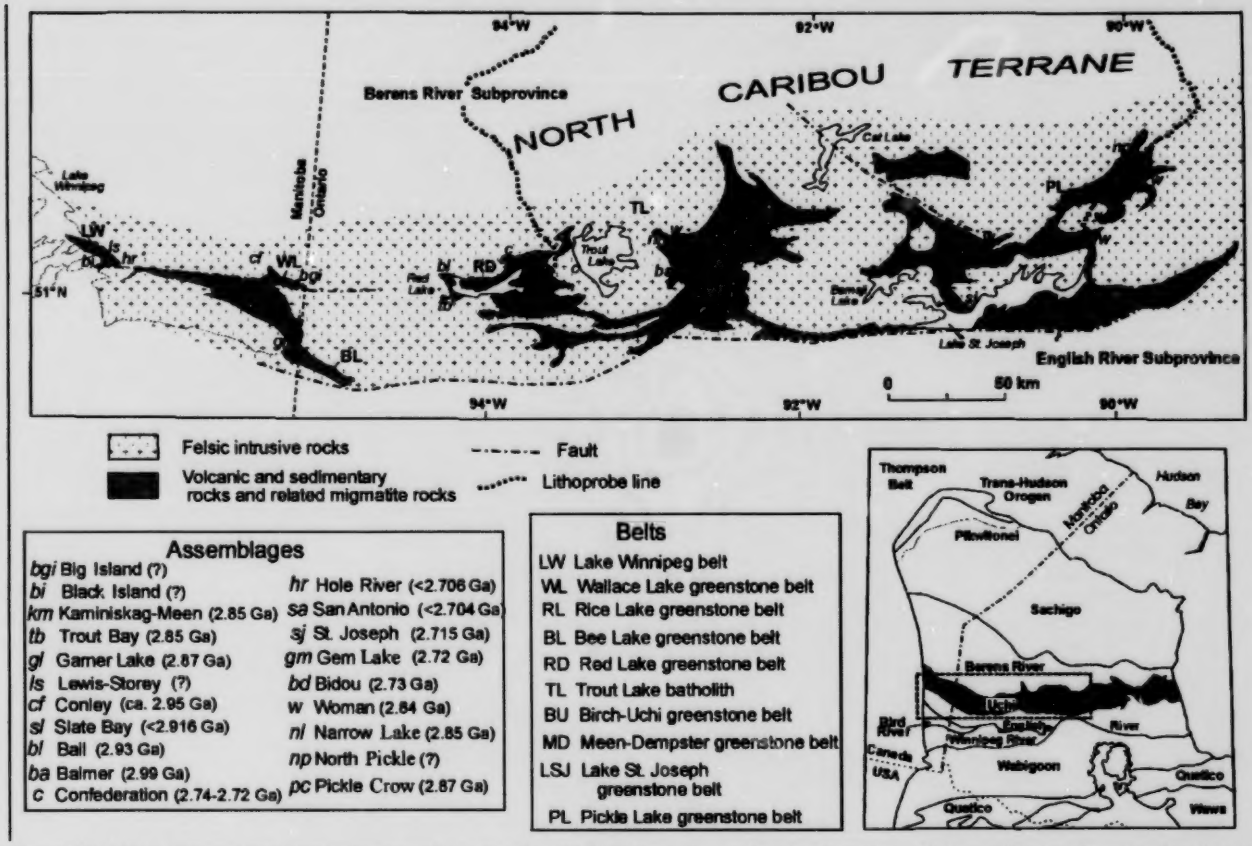

Figure GS-20-2: Generalized tectonic map of the western Uchi Subprovince, showing locations of assemblages discussed in the text (modified from Stott and Corfu, 1991; Sanborn-Barrie et al., 2000).

The East Shore complex and English Lake complex to the east (3006–2992 Ma) share distinctive geochemical features. Rocks ranging in composition from tonalite to diorite and gabbro have multi-element profiles with modest negative Ti and Nb anomalies, and strongly depleted Th and Rb. These primitive-arc characteristics may be the product of an ultradepleted mantle source.

Lewis–Storey assemblage

A west-facing sedimentary-volcanic sequence lies unconformably above the East Shore complex and may represent a Mesoarchean rift succession (Thurston and Chivers, 1990). A similar quartzite–carbonate–iron-formation–komatiite package has been dated at Wallace Lake, 70 km to the east, at between 2.92 and 3.0 Ga (Poulsen et al., 1996; Sasseville and Tomlinson, 2000; Tomlinson and McNicoll, unpublished data, 2000). The unconformity, documented at six localities over 30 km of strike length, separates tonalite from a unit of coarse, massive to thick-bedded grit with a chloritic matrix. Thinly laminated, fine-grained quartz arenite occurs in units up to a few metres thick, in association with muscovite-rich schist, aluminosilicate-bearing schist and talc schist. Overlying komatiite has spinifex with radiating clinopyroxene up to 20 cm long (*see* Percival and Whalen, 2000, Fig. 5). Associated iron-formation consists of centimetre-scale layers of carbonate, chert, magnetite and sericite schist. Mafic sills up to 15 m thick intrude both the basement and cover sequence adjacent to the unconformity.

Black Island sequence

The homoclinal, west-facing Black Island sequence comprises two basalt-dominated packages (Bailes and Percival, Fig. GS-26-5, this volume). A lower stack of thick massive flows, the Gray Point sequence, has flat multi-element profiles (approx. 5x chondrite) with light rare-earth element (LREE), Th and Nb depletion. It is separated by a thin volcaniclastic unit from an upper unit of pillow basalt, the Drumming Point sequence, of calc-alkaline affinity. A thin unit of lithic wacke caps the sequence. Ermanovics and Wanless (1983) reported an age of 2732 Ma for Black Island rhyodacite; however, the relationship between this east-facing outcrop, located structurally beneath the basalts, and the main Black Island sequence is not known.

Hole River sequence

Pink and grey arkose and rare conglomerate of the Hole River sequence occur mainly on islands. The fine- to coarse-grained sandstone contains detrital zircons with ages as young as 2706 Ma (V. McNicoll and J. Percival, unpublished data, 2000).

All units carry a penetrative, north-trending, steeply west-dipping S_1 foliation and widespread concordant D_1 high-strain zones up to 40 m wide. Shear-sense indicators, including late D_1 'Z' folds (F_{1a}), suggest dextral kinematic sense. In the eastern part of the area, ductile D_1 fabrics are truncated by sheets and plutons of biotite granodiorite that may be correlative with granodiorite dykes with a poorly constrained age of ca. 2715 Ma (Krogh et al., 1974). Southwest-trending S_2 crenulations are common in thinly laminated high-strain (D_1) zones. A steeply dipping, east-trending zone of dextral greenschist-facies high strain (D_3) forms the linear shoreline in the vicinity of Seymourville. It translates the basement-cover sequence approximately 7 km to the west, into the Pelican Harbour area.

A poorly preserved sedimentary-volcanic sequence lies along the southern margin of an east-trending body of hornblende-biotite tonalite in the Wanipigow River area. Both the tonalite and supracrustal rocks resemble those in eastern Lake Winnipeg, although an unconformity

was not observed. The tonalite is bordered to the south by a grit unit up to 10 m thick and is overlain by fine-grained, thinly laminated siliceous siltstone. Sporadic magnetite-chert and hematite-chert ironstone is also present at this structural level. Gabbroic sills with minor serpentinite schist layers occur structurally above the iron-formation, as well as within the tonalite.

A working hypothesis involves deposition of the Lewis–Storey rift sequence on the western margin of 3 Ga basement of the North Caribou terrane, with subsequent spreading recorded by mafic sill emplacement. The margin collapsed as the Black Island sequence was juxtaposed, probably in a strike-slip regime, ca. 2.72 Ga. Hole River sediments were deposited in strike-slip basins during continued D_1 oblique convergence. Waning deformation became focused in shear zones during D_3 and D_4 transcurrent shear events.

Bee Lake Belt

The Bee Lake greenstone belt (Fig. GS-20-1, -2), which straddles the Manitoba–Ontario border, represents the southeastern extension of the Rice Lake belt. Due to its limited (float plane only) access and relatively small size (approx. 170 km^2), this belt has not been mapped in detail since Shklanka (1967). Consequently, no geochronology or geochemical data are available.

The Bee Lake belt consists of a relatively simple volcano-sedimentary sequence that has undergone two ductile deformation (folding) events (referred to herein as D_1 and D_2), as well as subsequent brittle-ductile faulting. These rocks have been intruded by several generations of dykes and plutons, from which the timing of deformation events will be constrained.

The volcano-sedimentary sequence consists of a basal suite of probably calc-alkaline basaltic flows, pillow basalt and pillow breccia, which are overlain by interbedded felsic volcanic and clastic sedimentary rocks. The felsic volcanic rocks are predominantly epiclastic volcanic breccia, whereas the sedimentary rocks typically consist of arkosic sandstone and conglomerate. Locally, the conglomerate contains boulder-sized clasts of granodiorite that have undergone little or no pre-entrainment deformation.

The main deformation event in the belt, D_1, consists of at least three pulses of folding, interspersed with intrusions of tonalitic dykes that are generally transposed nearly parallel to the S_1 cleavage. The D_1 folds trend west-northwest, with a major east-southeast-plunging syncline centred on Bee Lake and a corresponding anticline centred on Odd Lake. The D_2 deformation is synchronous with intrusion of the Wingiskus Lake granodiorite pluton, which forms the core to a dome. The D_1 and D_2 structures are cut by late granodioritic intrusions, such as the Reahil Lake pluton. All rocks are cut by conjugate sets of brittle faults that are probably related to movement of the Sydney Lake–Lac St. Joseph Fault, which marks the southern margin of the belt.

The Bee Lake greenstone belt was previously correlated with the Confederation assemblage. However, the noticeably different rock types and stratigraphy between this region and the type section for the Confederation assemblage in the Birch–Uchi greenstone belt, and its extension in the Red Lake greenstone belt, suggest that this correlation may not be valid. Rather, the Bee Lake belt may correlate with the ca. 2.72 Ga Gem Lake subgroup of the adjacent Rice Lake greenstone belt.

Red Lake Greenstone Belt

The second field season of western Superior NATMAP activities in the Red Lake greenstone belt (Fig. GS-20-2; cf. Sanborn-Barrie et al., 2000) focused on integrating geological mapping with new and previously available lithogeochemical, geochronological and GIS data to further advance an understanding of 1) the interplay between magmatism, alteration, deformation and mineralization; 2) the relationships between Meso- and Neoarchean assemblages; and 3) the tectonic setting of the belt through 300 m.y. of Archean history. Collectively, these data highlight a history of episodic magmatic activity and sedimentation along the south margin of the North Caribou protocraton, and orogenic activity that culminated in collision with the Winnipeg River protocraton during a ca. 2.72 Ga phase of the Kenoran orogeny.

The earliest magmatism is represented by the ca. 2.99 Ga Balmer assemblage, which is dominated by submarine, high-TiO_2 tholeiitic basalts that range from LREE and large-ion lithophile element (LILE) depleted, to slightly enriched with small positive Th/Nb anomalies. Komatiite (olivine spinifex) and komatiitic basalt (pyroxene spinifex) are LREE depleted with positive Th/Nb anomalies. In contrast, komatiite and basalt of the ca. 2.94 to 2.92 Ga Ball assemblage are associated with voluminous intermediate to felsic volcanic rocks, have calc-alkaline affinities and low TiO_2 contents, and show a greater degree of LREE and LILE enrichment. Resurgence of calc-alkaline magmatism and volcaniclastic deposition at 2894 Ma, followed by quiescent marine conditions, is recorded by the Bruce Channel assemblage. New field, geochronological and geochemical data from the southwest part of the belt indicate the presence of the Trout Bay assemblage (Fig. GS-20-1), consisting of ca. 2.85 Ga intermediate tuff and marine siliciclastic and chemical sedimentary rocks overlain by coarse fragmental rocks and submarine, low-TiO_2 tholeiitic basalt that are strongly depleted in LREE and LILE. Finally, new data allow subdivision of the Neoarchean Confederation assemblage into the 2748 to 2742 Ma McNeely sequence, a shallow marine, calc-alkaline, mafic to felsic succession; the overlying 2744 to 2739 Ma Heyson sequence of tholeiitic basalt, F3 rhyolite and minor calc-alkaline volcanic rocks; and the younger ca. 2.73 Ga Graves sequence, comprising an intermediate, calc-alkaline, volcano-plutonic complex.

This protracted history of magmatic activity is punctuated by sedimentary episodes involving deposition of locally derived clastic detritus and precipitation of chemical sediments. Compositionally immature clastic rocks occur within the Balmer, Bruce Channel and Trout Bay assemblages, in contrast to compositionally mature fuchsitic quartz arenite and conglomerate interstratified with ca. 2.94 to 2.925 Ga felsic volcanic rocks of the Ball assemblage and forming the post-2912 Ma Slate Bay assemblage that disconformably overlies Balmer basalt. A regionally extensive unit of polymictic conglomerate marks the interface between Mesoarchean and Neoarchean strata, consistent with an emerging view that the main stages of crustal growth in the Red Lake belt from ca. 3 Ga to 2.7 Ga were primarily depositional, not tectonic.

The Red Lake greenstone belt is polydeformed, with an early (pre-2748 Ma) nonpenetrative deformational event (D_0) and at least two recognizable generations of ductile structures (D_1, D_2) imposed after ca. 2742 Ma volcanism. Early deformation involved overturning of Balmer pillowed basalt, as documented by opposing facing directions on either side of the regionally extensive angular unconformity in the Madsen and central Red Lake areas, and is interpreted to have involved recumbent folding. The main stages of penetrative deformation resulted in two sets of folds (F_1 and F_2), established by opposing younging and structural-facing directions. Planar and/or linear fabrics associated with D_1 trend northerly and are locally recognized throughout the belt, whereas widespread, weakly to moderately developed L-S fabrics associated with F_2 folds (D_2) trend east to northeast. The timing of deformation is established by field relationships involving the 2718 Ma Dome Stock, which contains foliated xenoliths of local country rock but is itself weakly to moderately foliated with a penetrative northeast-striking fabric, characteristic of the regional S_2 foliation. The orientation of interpreted D_2 fabrics deviates from the regional (east- to northeast-striking) S_2 trend to an east-southeast–striking, low-strain corridor between Cochenour and Balmertown. This deviation may have evolved through shifting local boundary conditions in the eastern Red Lake belt if, as predicted, the Trout Lake Batholith (Fig. GS-20-1) includes a major phase of ca.

2.72 Ga magmatism. Localization of strain into a belt-scale system of conjugate shear zones (cf. Andrews et al., 1986) was not substantiated during this study.

The majority of gold deposits in the Red Lake district are the end-product of hydrothermal replacement and constitute atypical greenstone-hosted gold types. The Campbell–Red Lake deposit consists mainly of early to syn-D_2, quartz- and arsenopyrite-rich, selective replacement zones of colloform-crustiform iron-carbonate veins and breccia. These silicified carbonate veins are hosted mainly by heterogeneously foliated basalt within the east-southeast-trending D_2 strain corridor informally designated the 'mine trend'. The Madsen deposit is an amphibolite-grade, disseminated-replacement–style gold deposit hosted by mafic volcaniclastic rocks. On the basis of style and mineralogy, this deposit has been interpreted as a gold-rich skarn-like deposit (Dubé et al., 2000). All the major Red Lake deposits are proximal to a regionally extensive angular unconformity between the ca. 2.99 Ga Balmer and ca. 2.74 Ga Confederation assemblages.

The district has been affected by large-scale hydrothermal alteration systems, resulting in strong to intense calcite carbonatization that affects rocks of the Balmer, Ball, Bruce Channel and Confederation assemblages. Less extensive areas of strong to intense ferroan-dolomite alteration enclose the majority of gold deposits in the greenstone belt and represent proximal alteration envelopes that affect the Balmer, Ball and Confederation assemblages. The proximal zones are characterized by a greater volume of ferroan-carbonate veins and disseminated sulphide mineralization than in calcite-altered areas. Potassic alteration, in the form of pervasive biotitization, occurs in proximal alteration zones. Metamorphosed proximal alteration zones are characterized by a diverse assemblage of aluminosilicate minerals such as andalusite, staurolite, garnet and chloritoid.

Carbonatized mafic metavolcanic rocks adjacent to the Killala–Baird Batholith have been metasomatized and pervasively replaced by a calc-silicate assemblage of diopside, orange garnet, epidote, tremolite, calcite and quartz. Diopside veins also cut and replace early ferroan-carbonate veins and are associated with gold zones at Madsen. If it can be shown that diopsidic veins and skarn development are related to a phase of the Killala–Baird Batholith, the batholith could be implicated in gold mineralization of this part of the belt. This would, in turn, open up new areas of exploration for skarn-type gold deposits in the camp.

The tectonic evolution of the Red Lake belt appears to have involved ca. 2.99 Ga plume-related rifting, followed by development of a ca. 2.94 to 2.92 Ga calc-alkaline arc complex (cf. Tomlinson et al., 1998; Hollings et al., 1999). Intercalation of komatiite units within this arc complex may reflect rifting and/or impingement of a plume in a subduction-zone setting (Hollings et al., 1999). Renewed arc volcanism at 2.89 Ga, followed by tectonic quiescence and subsidence across the continental margin, may reflect major changes in relative plate convergence, in which orthogonal convergence was accompanied by arc volcanism and oblique convergence by magmatic quiescence. This may have led to accretion of the 2.85 Ga Trout Bay island arc–back arc complex with the southern North Caribou margin, resulting in pre-2.75 Ga recumbent folding, uplift and sedimentation. Unconformably overlying rocks record a Neoarchean history of renewed emergent-arc volcanism at 2.75 to 2.74 Ga (McNeely sequence), followed by subsidence and intra-arc rifting at 2.74 Ga (Heyson sequence). A final phase of Andean-style arc magmatism at ca. 2.73 Ga, recorded at Red Lake (Graves sequence) and throughout the Berens arc, is diachronous along the margin between 2.73 and 2.72 Ga. This is interpreted to reflect subduction beneath the southern North Caribou margin, leading to collision with the Winnipeg River protocraton at ca. 2718 Ma, during the Uchian phase of the Kenoran orogeny.

A co-operative GIS-database project involving the Geological Survey of Canada (GSC), the Ontario Geological Survey (OGS) and mineral industry partners is underway in the Red Lake belt. The purpose is to assemble, integrate and release hard-copy maps and a CD-ROM containing all available digital and analog geoscience data. To date, emphasis has been on compiling public-domain information. The main geoscience themes compiled to date include bedrock and surficial geology; regional lithogeochemistry; diamond drilling and other types of mineral exploration assessment work; airborne magnetic, EM and gamma-ray data; and Landsat imagery. A preliminary release of data in ArcInfo, ArcView and MapInfo formats is scheduled for early December. A short field program was conducted in June 2000 to sample areas of anomalous carbonate, sericite and silica alteration, based on preliminary analysis of lithogeochemical data. Analyses will be compared with the regional data set to assess possible patterns. The data will be integrated with other exploration tools and criteria (regional unconformities, fault zones, etc.) in the GIS to produce mineral-prospectivity maps.

Confederation–Birch–Uchi Belt

Additional geochronology is currently being conducted in the Confederation–Birch–Uchi greenstone belt (Fig. GS-20-2) in an attempt to constrain whether the ca. 2.84 Ga Woman assemblage is present at its type locality. Initial attempts to directly date this sequence failed, whereas chemically and petrographically indistinguishable rocks adjacent to the type locality have yielded dates typical of the Confederation assemblage (ca. 2.74 Ga). However, evidence for a pre-Confederation assemblage has been obtained in the form of a ca. 2.81 Ga quartz-feldspar porphyry dyke along the western shore of Woman Lake (V. McNicoll, unpublished data, 2000). Hence, the likely volcanic stratigraphy for the Birch–Uchi greenstone belt is as follows:

1) Balmer assemblage: andesitic to rhyolitic volcanism (2989–2975 Ma), separated by an early (D_0) deformation event from
2) Narrow Lake assemblage: tholeiitic pillow basalt that unconformably overlies the Balmer assemblage (bracketed between 2975 and 2832 Ma; most likely ca. 2855 Ma from comparison to the Red Lake greenstone belt)
3) Woman assemblage: pillow basalt overlain by ignimbritic rhyolite that disconformably overlies the Narrow Lake assemblage (bracketed between 2.88 and 2.81 Ga)
4) Confederation assemblage: three petrographically, chemically and spatially distinct belts of mafic to felsic volcanism (ca. 2.74 Ga)

Pickle Lake Belt

Fieldwork in the Pickle Lake area (Fig. GS-20-2) during the 2000 field season was aimed at testing the hypothesis that the lithotectonic assemblages constituting the Pickle Lake greenstone belt were either built on, or accreted to, the south margin of the ca. 3.0 Ga North Caribou Terrane protocraton. Stott's (1996) summary of stratigraphic, structural and geochronological studies favoured an accretionary model for the southern part of the Pickle Lake belt. However, new results to the west (Birch–Uchi and Red Lake greenstone belts; Rogers et al., 2000; Sanborn-Barrie et al., 2000) have established that the main stages of crustal growth were primarily depositional, not tectonic.

Of the four lithotectonic assemblages making up the Pickle Lake belt, concerted effort was directed at the two northerly assemblages, the North Pickle and Pickle Crow. The North Pickle assemblage is composed mainly of massive and pillowed basalt flows, interbedded with carbonate-siliceous banded iron-formation and minor clastic units, as well as sill-like mafic intrusive sheets. The assemblage was interpreted as the southern extension of the ca. 3.0 Ga McGruer assemblage of the North Caribou belt, based on aeromagnetic patterns (Stott, 1996). However, limited facing directions determined from pillow tops indicate a northwesterly younging direction, opposite to that of the McGruer assemblage. The Pickle Crow assemblage consists of massive to pillowed basaltic flows, dacitic pyroclastic flows, carbonate-siliceous

banded iron-formation interbeds (which are macroscopically very similar to those of the North Pickle assemblage), and a locally derived polymictic conglomerate where most clasts are of the same provenance. Limited facing directions, determined from grading in the conglomerate and sparse pillow tops, indicate local northwesterly younging, but tight folds preclude attaching regional significance to this determination. Quartz porphyry sills intruded the basaltic sequence and zircons from one sill yielded a crystallization age of 2.86 Ga and an inherited age of 2892 Ma (Corfu and Stott, 1993).

Two major deformation events and an early localized strain event are recorded in the two northerly assemblages. The main regional event, D_1, is defined by a belt-wide, weakly to moderately developed, northwest-dipping foliation generally parallel to stratigraphic contacts, with similar trends in both assemblages. In the North Pickle and southern Pickle Crow assemblages, S_1 fabric is weakly developed. It is more prominent in the north-central Pickle Crow area, in a zone of tight F_1 folding. The D_1 fabric is overprinted by contact strain aureoles, here termed D_2, associated with ca. 2740 Ma felsic plutons and stocks. The intensity of D_2 is variable across the belt, ranging from a moderate crenulation in the east to strong, amphibolite-grade flattening in the west. Evidence of pre-D_1 deformation is found in the north-central Pickle Crow area, where D_1 fabrics related to F_1 folds crenulate an older foliation developed in carbonatized basalt. The pre-D_1 fabric is cut by ca. 2.86 Ga quartz porphyry sills, indicating localized Mesoarchean deformation. The aeromagnetic pattern shows that stratigraphic and fold-axis trends are similar in both assemblages across the assumed boundary. Also, the boundary itself is at an angle (approx. 30° clockwise) to stratigraphic contacts, and fold axes are not parallel to the fabric in the Pickle Crow, as would be predicted if the Pickle Crow assemblage had accreted onto the southern margin of the North Pickle assemblage.

Without absolute age constraints and in light of the recent observations, evidence is insufficient to separate the North Pickle and Pickle Crow assemblages into distinct lithotectonic entities. The aeromagnetic discordance between the two appears to be related to a fault that offsets sequences of the same lithotectonic assemblage. Questions remain as to the age and kinematic significance of the aeromagnetically defined fault.

WABIGOON SUBPROVINCE

Onaman–Tashota Greenstone Belt

The Onaman–Tashota greenstone belt forms the central feature of the eastern Wabigoon Subprovince. The main portion of this belt has been remapped in the past 5 years (Fig. GS-20-1), and the results show a much greater volume of synvolcanic to late orogenic gabbroic intrusions than previously thought (e.g. Stott and Straub, 1998). Current geochronological (D. Davis and colleagues, Royal Ontario Museum), Nd isotopic, and geochemical work on widely distributed rock samples from this belt will be used to assist in subdividing the belt into several tectonostratigraphic assemblages. Some of these assemblages are, in a first-order sense, apparent from significant differences in age of volcanism. Analysis of the chemistry and petrogenetic characterization of assemblages is currently in progress; some age determinations remain to be completed in areas where intermingling of assemblages is being resolved. The current geochronological record, including the most recent U-Pb zircon age determinations from various units in the Onaman–Tashota belt, shows a more varied history of volcanism than previously thought. On the basis of field relationships, some of the ages from intrusions are interpreted to be inherited, reflecting older, underlying crust. A spectrum of Meso- to Neoarchean ages is represented in detrital zircons from sandstone units across the belt, including the southern margin of the English River Subprovince.

Chondrite-normalized rare-earth element (REE) patterns for almost 200 samples of tholeiitic basalt across the width of this greenstone belt are mostly flat or have weakly depleted LREE and slightly greater than 10x chondrite. Some units show a slight negative Eu anomaly. Trace-element variation diagrams for basalt samples suggest a predominantly juvenile, non-arc affinity, with exceptions in the southern part of the belt underlying tuff sequences of the 2.74 Ga calc-alkaline volcanic centres and some strata in the vicinity of Willet Lake, east of Toronto Lake.

Built upon these assemblages of basalt are calc-alkaline felsic to intermediate volcanic centres that decrease in age southward from the 2739 Ma Marshall Lake assemblage, through 2734 to 2722 Ma volcanic centres at Venus Lake and Metcalfe Lake, respectively, to a less than 2707 Ma dacitic sequence on Humboldt Bay, Lake Nipigon. The latter might correspond episodically with less than 2707 Ma metasedimentary strata above the Metcalfe Lake dacite and a 2703 Ma dacitic sequence west of Lake Nipigon (Tomlinson et al., 1999, 2000) in the northern Obonga belt. The intermediate to felsic volcanic rocks across the Onaman–Tashota belt are typically strongly enriched in Th and have a pronounced negative Nb anomaly. They show LREE enrichment and heavy REE depletion with little or no Eu anomaly, features that are typical of calc-alkaline arc magmatism. The presence of large, 2735 to 2740 Ma, felsic volcanic centres on the north and south margins of the greenstone belt creates an apparent symmetry of crustal growth on the flanks of the belt. These broad centres contain synvolcanic plutons and accompanying hydrothermal alteration, and the most notable base-metal mineralization in the belt (e.g. Marshall Lake). Synorogenic alteration associated with gold mineralization is concentrated in three domains: the complexly faulted Beardmore–Geraldton belt, strained zones in the Metcalfe Lake felsic volcanic centre, and farther north in the Tashota Lake–Emily Lake road area.

A suite of 64 samples of volcanic and intrusive rocks from across the greenstone belt has been analyzed for Nd isotopes. The felsic rocks (volcanic and plutonic) from throughout the region show predominant Nd model ages of 2.8 to 2.9 Ga, and initial epsilon Nd (ε_{Nd}) values of -0.1 to +1.5 (28 of 41 samples). Five samples from the north, at Marshall Lake, Toronto Lake, Girvan Lake, Tashota and the Esnagami tonalite, have Nd model ages in the 3.0 to 3.3 Ga range. Both the synvolcanic Onaman pluton at the centre of the belt and the late tectonic Sollas pluton on the northern margin have Nd model ages of 2.9 to 3.0 Ga. Several other synvolcanic to late tectonic plutons from throughout the region have Nd model ages of 2.8 to 2.7 Ga. The mafic rocks have initial ε_{Nd} values of +1.0 to +4.0, with the majority between +2 and +3, indicating a lack of significantly older recycled crust. The exception is at Toronto Lake in the north, where lower ε_{Nd} values are observed.

Together, the isotopic data suggest several conclusions. First, 2.74 to 2.72 Ga felsic volcanism cannibalized older felsic crust. Widespread evidence of recycled early Neoarchean to Mesoarchean crust is comparable to the complex pattern of crustal contamination noted by Tomlinson et al. (1999, 2000) in the central Wabigoon. Second, associated mafic volcanism was relatively juvenile. Third, the basement in the northern half of the belt may contain a much older crustal component than is currently recognized at the surface. The oldest dated rock from this region, a porphyry sill, is 3056 Ma, but three felsic volcanic rocks have significantly older Nd model ages of 3.28, 3.21 and 3.19 Ga.

The overall structural geometry of the belt has a remarkably consistent northward dip. Bedding and parallel tectonic schistosity dip moderately to steeply north, similar to those in the Shebandowan greenstone belt in the Wawa Subprovince and the central part of the Uchi Subprovince. This regional structural observation across widely spaced greenstone belts may mirror the dominant, north-dipping, crustal reflectivity from Wawa to Uchi subprovinces, evident from the Western Superior LITHOPROBE Transect (White et al., 1999). This structural character continues across the northern boundary of the Onaman–Tashota belt into the English River Subprovince. The northern margin of the Onaman–Tashota belt is affected by regional D_2 transpressive deformation. Northeast of Marshall Lake at the English River Subprovince boundary, a 2692 Ma intrusion, possibly a phase of the Sollas pluton, is interpreted, from its internal structures and relation to

surrounding metasedimentary rocks, to have intruded during the late stage of regional D_2 deformation. This deformation corresponds in style to that observed throughout the English River Subprovince (Stott and Corfu, 1991; Corfu et al., 1995) and adjacent greenstone belts. It significantly predates a similar, transpressive deformation that affected the Quetico Subprovince and adjacent greenstone belts at 2685 to 2680 Ma (Corfu and Stott, 1998).

South-Central Wabigoon

A project aimed at defining the extent of Mesoarchean crustal precursors in the south-central Wabigoon Subprovince stems from recent mapping and isotopic studies that indicate scattered material of 3.4 to 3.0 Ga age (Henry et al., 1998; Tomlinson et al., 1999; Percival et al., 1999; Tomlinson, 2000) in this region. In the Shikag–Garden Lake area, tonalite, tonalite gneiss and anorthositic rocks generally have Nd model ages of less than 2.9 Ga, whereas granodiorite and granite have model ages in excess of 3.0 Ga and, in two locations, in excess of 3.3 Ga (Tomlinson and Percival, 2000). Farther to the west, biotite granite of the Indian Lake Batholith (north of Ignace) has Nd model ages, from two locations, of 2.80 and 2.84 Ga, and older precursors are indicated in tonalite gneiss of the Hillyer Creek dome (2.95 Ga), felsic fragmental rocks of the Perch Lake belt (3.03 Ga), tonalite gneiss west of Mine Centre (2.98 Ga), tonalite gneiss northwest of Eltrut Lake (2.90 Ga), tonalite gneiss from the Valora area (2.91 Ga) and foliated tonalite south of Ignace (2.85 Ga; Tomlinson and Percival, 2000). A significantly older Nd model age of 3.26 Ga comes from a hornblende-biotite tonalite gneiss west of Raven Lake. Collectively, these data suggest that Mesoarchean crust in the southern Wabigoon Subprovince is not restricted to the Marmion Batholith and the Steep Rock and Lumby Lake greenstone belts, but appears to extend west, at least to Mine Centre in the southwest and Valora in the northwest, and northeast to the Obonga Lake–Garden Lake area. Model ages greater than 3.34 Ga suggest the presence of crust of Winnipeg River vintage beneath some parts of the central Wabigoon (Tomlinson and Percival, 2000). The U-Pb geochronology will further constrain the age and extent of crustal precursors.

Mapping of the Petry River area in the south-central Wabigoon Subprovince focused on establishing regional controls on gold, platinum-group element and rare-earth element mineralization, as well as examining the origin of and relationships between Mesoarchean and Neoarchean rocks. Within this area, narrow Archean greenstone belts are cut by several varieties of foliated to gneissic plutonic rocks and massive granite batholiths. The greenstone belts comprise metabasalt with rare gabbro, sedimentary units and a single komatiite locality. A platform sequence, represented by a 30 m thick section of quartz arenite, separates metabasalt from biotite tonalite in an area 25 km northwest of Graham, Ontario. The platform sequence and the komatiite provide geological evidence that volcano-plutonic sequences of the Petry River area form an extension of the Mesoarchean domain in the Atikokan–Lumby Lake area.

Regional mapping traced the Marmion Fault 100 km northeast from Atikokan to Selwyn Lake. Splays of the fault are mineralized with gold in the Atikokan area and dykes of peraluminous granite are broadly associated with the fault at Selwyn Lake. The peraluminous granite dykes, some of which contain visible molybdenite, are a possible bedrock source for anomalous concentrations of Li, Mo, Th, U, rare-earth elements, Y and Nb in lake-bottom sediment in the Selwyn–Wawang lakes area. A mafic intrusive complex of Archean gabbro and Proterozoic diabase occurs at Little Trewartha Lake, and a hornblendite stock is present south of Weaver Lake. These intrusions are recommended as exploration targets for platinum-group elements.

SACHIGO SUBPROVINCE

Blackbear, Yelling and Stull Lakes Areas

Mapping (1995–2000) of a 12 000 km^2 area along the Ontario–Manitoba border (Fig. GS-20-1) is providing new geological maps and an assessment of the mineral potential and regional tectonic evolution. The east-southeast-trending Ponask–Sachigo, Stull, Ellard and Yelling greenstone belts are separated by plutonic domains and the broad Stull–Wunnummin, South Kenyon and North Kenyon fault zones. Work in the current season at Blackbear Lake marked the completion of regional mapping in the southeast corner of the area. Eastern extensions of the Stull and Ellard greenstone belts, as well as the Stull–Wunnummin and South Kenyon faults, were traced through this area. Transects along the Stull, Wapikani and Yelling rivers identified western extensions of major batholiths, greenstone slivers and the Kenyon faults. The largest greenstone belt in the area, Stull Lake, consists of four structural panels (Stone and Pufahl, 1995), and has been re-examined in detail in the last two seasons. New mapping, sampling and geochronology have focused on defining positions of major faults in the belt, examining its poorly exposed western extremities, distinguishing alkaline and calc-alkaline volcanic sequences, and studying the relationships between various sedimentary and volcanic packages.

Preliminary isotopic and geochronological studies indicate that the area is divisible into an early Neoarchean to Mesoarchean domain south of the Stull–Wunnummin Fault, a Neoarchean domain between the Stull–Wunnummin and Kenyon faults, and a Paleoarchean domain north of the Kenyon Fault (Skulski et al., 1999). These domains possibly represent crustal blocks that were tectonically accreted during growth of the Superior Province and whose boundaries have been subsequently reactivated as faults. A lack of regional data inhibits extrapolation of these domains across the northern Superior Province or correlation with other tectonic domains in the south.

The northern Superior Province has potential for a wide variety of mineral commodities, including rare-earth elements, platinum-group elements, base metals, gold, diamonds and carving stone. Numerous gold occurrences associated with regionally extensive faults indicate high potential for economic deposits of gold. Anomalous numbers of kimberlite-indicator grains in modern alluvium suggest that the area is worthy of exploration for diamonds.

Knee Lake–Gods Lake Area

Integrated studies of the Knee Lake belt (cf. Corkery et al., 1999) continued in the northeastern Gods Lake area in 2000 (Fig. GS-20-1). Two distinct sequences of Hayes River Group metavolcanic rocks with opposing facing directions are separated by a northeast-trending dextral fault. In addition to its transcurrent component, north-side-up dip-slip motion on the fault carries amphibolite-facies rocks on the northwest over greenschist-facies units to the southeast. Geochronological studies on units of the Oxford Lake Group to the southwest show that the volcanic subgroup has a depositional age of 2720 Ma, whereas the sedimentary subgroup has a broad range of detrital zircon ages ranging from 3647 to 2711 Ma. Possible sources of ancient (ca. 3.6 Ga) zircons occur in fault-bounded blocks along the northern margin of the northern Superior Province. Geochronological data from the Knee Lake area date Hayes River volcanism at 2.83 Ga, and sensitive high-resolution ion microprobe (SHRIMP) data on detrital zircons indicate that turbiditic sandstone is synvolcanic and volcanic-derived. Folding of the Hayes River Group predates deposition of coarsening-upward clastic sedimentary rocks and iron-formation, the former containing 2937 to 2822 Ma detrital zircons. The Oxford Lake volcanic subgroup was extruded at 2722 Ma in the southern Knee Lake area and was followed by deposition of fluvial-clastic, sedimentary subgroup sandstone with detrital zircons ranging from 2798 to 2707 Ma.

A unique feature of Gods Lake is that it provides access to the granitoid terrane separating the Munro Lake and Knee Lake greenstone belts (Fig. GS-22-1). This well exposed section of gneiss complexes and plutons was mapped and sampled for geochemistry in a co-operative MGS–GSC program. As well as contributing to the Knee Lake greenstone belt mapping project, this section will provide geochemical,

isotopic and geochronological information to a regional transect focusing on the age and history of plutonic and gneissic terranes across the northwestern Superior Province.

Assean Lake Area

The Assean Lake block at the northwestern margin of the Superior Province represents a collage of east-southeast-trending Archean crustal segments that is overprinted by northeast-trending zones of Neoarchean and Paleoproterozoic deformation and metamorphism. The Assean Lake block can be subdivided into three lithotectonic domains. In the south, a belt of migmatitic supracrustal rocks includes quartz arenite, arkose and metagreywacke, with some amphibolite and silicate-facies iron-formation. A central panel is dominated by tonalitic orthogneiss, which cuts the northern package consisting of mafic to intermediate metavolcanic rocks and migmatitic greywacke.

Combined Sm-Nd isotopic and U-Pb zircon (thermal-ionization mass spectrometry (TIMS) and SHRIMP) results indicate that the Assean Lake block preserves Paleo- to Mesoarchean crust that underwent a complex and prolonged history spanning more than two billion years. These integrated mapping and isotopic studies across the northwestern Superior Province margin have resulted in a re-interpretation of the location and nature of the boundary zone between Archean rocks of the Superior Province and Paleoproterozoic rocks of the Trans-Hudson Orogen in the region northeast of Thompson.

Northern Transect

The poorly exposed Northern Superior Superterrane (Skulski et al., 2000) was mapped and sampled for geochemical, tracer isotopic and geochronological studies in a transect from Kistigan Lake to northeast of Red Cross Lake. Major geological domains sampled include the northern Sachigo Subprovince and Pikwitonei gneiss complex. The northwest-trending North Kenyon Fault is a kilometre-scale mylonite zone; narrower mylonite zones of similar trend characterize the South Kenyon Fault.

DISCUSSION

Elucidation of the tectonic history of the Uchi Subprovince remains key to understanding the relationship between the North Caribou and adjacent terranes. There exists widespread yet cryptic evidence of an early, nonpenetrative deformation event (pre-D_1) across the southern margin of the North Caribou craton, in the Pickle Lake (pre-2860 Ma), Confederation Lake (ca. 2.85 Ga) and Red Lake (2894–2750 Ma) greenstone belts. The first penetrative deformation event (D_1) recorded in this region may have been diachronous from pre-2.74 Ga in the Pickle Lake area, to between 2733 and 2724 Ma in the Confederation Lake area, to between 2742 and 2718 Ma in the Red Lake belt. The timing of subsequent D_2 deformation, interpreted to be related to collision between the North Caribou and Winnipeg River cratons (Uchian phase of the Kenoran orogeny) appears best constrained in the Red Lake belt at ca. 2718 Ma. Diachroneity may also characterize this event, as units as young as 2713 Ma were affected in the St. Joseph assemblage to the west.

The southwestern part of the Uchi Subprovince (Rice Lake–Lake Winnipeg belts) shows little evidence of affinity with the North Caribou margin prior to 2.72 Ga, and records a younger tectonic history that involves tectonic juxtaposition of units after 2704 Ma, the age of post-tectonic plutons in the Red Lake and Confederation Lake belts. Post-2704 Ma events may be related to deformation that affected the English River Subprovince (cf. Corfu et al., 1995). Together, these observations show that the southern margin of the North Caribou Terrane had an extended history of magmatism and deformation, perhaps reflecting transient conditions at a long-standing Andean margin.

Within the Uchi Subprovince, evidence for the presence of crustal fragments that are exotic to the North Caribou margin is limited. Candidates include the Pickle Crow assemblage, with evidence of pre-2.86 Ga deformation, which would have been accreted prior to 2745 Ma. Geochemical, isotopic and geochronological studies are underway to test this hypothesis. The ca. 2.85 Ga Trout Bay assemblage in the Red Lake belt may represent a juvenile oceanic terrane that faces toward older strata built on the North Caribou margin and was juxtaposed prior to 2.75 Ga. The undated Black Island assemblage in the Lake Winnipeg area has ocean-floor and arc geochemical signatures, faces toward a continental-margin sequence, and is separated from it by D_1 shear zones formed after 2704 Ma.

Along the northern margin of the North Caribou Terrane, slivers of juvenile crust are bounded by major, late, strike-slip shear zones. Ongoing mapping, in concert with geochronology and isotopic-tracer studies, is unravelling the complex interaction between the North Caribou Terrane and Northern Superior Superterrane.

Ongoing work in the central and eastern Wabigoon regions is defining the age and extent of Mesoarchean crustal blocks and their relationships to Neoarchean sequences. At least two discrete Mesoarchean sources are present: 1) a northern, terrane, greater than 3.34 Ga, with probable affinity to the Winnipeg River Subprovince, known from ancient detrital zircons in Mesoarchean platform sequences and old Nd model ages in Neoarchean granite; and 2) a southern block, ca. 3 Ga, exposed in the Marmion Batholith and environs. Whether the two Mesoarchean blocks were assembled during tectonism ca. 2.7 Ga, which also juxtaposed juvenile Neoarchean rocks of the western Wabigoon, or carry an older common history, is an avenue of current research. Integrating the tectonic and magmatic history with crustal profiles derived from LITHOPROBE seismic data represents a unique challenge and opportunity.

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GS-21

GEOLOGICAL AND GEOCHRONOLOGICAL INVESTIGATIONS IN THE STULL LAKE-EDMUND LAKE GREENSTONE BELT AND GRANITOID ROCKS OF THE NORTHWESTERN SUPERIOR PROVINCE

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Skulski, T., Corkery, M.T., Stone, D., Whalen J.B. and Stern, R.A. 2000: Geological and geochronological investigations in the Stull Lake-Edmund Lake greenstone belt and granitoid rocks of the northwestern Superior Province; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 117-128.

SUMMARY

Field and isotopic data on plutonic and supracrustal rocks from the northwestern Superior Province reveal three, fault-bounded crustal terranes that record the collision between reworked 3.6 Ga crust in the north and a less than 3 Ga protocraton at 2.72 to 2.71 Ga. In the south, the Munro Lake terrane comprises less than 2.86 Ga rift-margin sedimentary rocks and komatiite built on the reworked margin of the 3 Ga North Caribou terrane. Plutonic rocks at 2.84 and 2.72 Ga have 3 Ga Nd model ages, reflecting recycling of the North Caribou margin. To the north, in the Oxford Lake–Stull Lake terrane, ca. 2.83 Ga, submarine, depleted tholeiitic basalt was thrust (D_1), prior to ca. 2.73 Ga, onto the Munro Lake terrane. A 2.73 to 2.72 Ga, isotopically juvenile, continental-margin arc was built on the accreted crust and is cut by 2.72 to 2.71 Ga, early to syntectonic, isotopically juvenile plutons. To the north, across the northwest-trending North Kenyon Fault (a 1 to 2 km wide dextral strike-slip greenschist mylonite), 2.84 to 2.71 Ga plutonic rocks of the Northern Superior superterrane have 3.6 to 2.9 Ga Nd model ages and zircon inheritance to 3.57 Ga. Docking of this reworked Paleoarchean crust with the Oxford Lake–Stull Lake terrane resulted in 2.73 to 2.72 Ga arc volcanism, oblique convergence (D_2-D_3) and metamorphism at less than 2.72 Ga. Eruption of synorogenic, less than 2.71 Ga, alkaline and shoshonitic lavas with negative epsilon Nd (ε_{Nd}) values, and later deposition of continental sediments with local 3.6 Ga detrital zircons, reflects underthrusting and uplift of older continental crust across the Oxford Lake–Stull Lake terrane.

INTRODUCTION

Archean crust in the Superior Province of northern Manitoba and Ontario is tectonically segmented on a number of scales. Major crustal blocks are defined by northwest-trending, dextral-slip greenschist-facies shear zones, spaced 10 to 50 km apart. Granitoid rocks dominate these crustal domains and provide windows to understanding the crustal evolution and deformation history in these blocks. Supracrustal rocks preserved in greenstone belts are, in part, cut by the major shear zones, but also preserve an internal tectonic subdivision that predates shear-zone formation. Unravelling the volcanic and sedimentary record using field, geochronological and geochemical methods provides a valuable perspective on the early tectonic history of this area, and the fate of its mineral and metal endowment.

Geological mapping since 1995 (integrated within Western Superior NATMAP) provides 1:50 000 scale control along a northerly trending, across-strike transect along the northern Ontario–Manitoba border between Sachigo and Yelling lakes (Fig. GS-21-1; Stone and Pufahl, 1995; Stone et al., 1996; Stone and Halle, 1997; Stone et al., 1999a; Stone et al., 1999b). These results, plus a Nd isotopic and U-Pb (by sensitive high-resolution ion microprobe [SHRIMP]) isotopic survey of plutonic rocks between Sachigo Lake and Yelling Lake, reveal that fault-bounded, isotopically juvenile, 2.71 Ga crust separates the 3.0 Ga northern North Caribou margin in the south from the greater than 3 Ga Northern Superior superterrane in the north. Recent, detailed (1:20 000 scale) mapping of the Stull Lake–Edmund Lake (Fig. GS-21-2) and Knee Lake greenstone belts provides key structural, stratigraphic and geochronological constraints on the broad collision zone between these two protocratonic blocks (Corkery, 1996; Corkery et al., 1997; Corkery and Skulski, 1998; Corkery and Heaman, 1998; Corkery et al., 1999; Syme et al., 1997, 1998). These data are consistent with a three-fold subdivision of the northwestern Superior Province in the map area (Fig. GS-21-1): 1) Munro Lake terrane (*modified after* Thurston et al., 1991), the crustal block lying south of a splay of the Wolf Bay–Stull–Wunnumman Shear Zone in the southern Stull Lake–Edmund Lake greenstone belt and including the Sachigo Lake–Ponask Lake greenstone belt; 2) Oxford Lake–Stull Lake terrane (*modified after* Thurston et al., 1991), the area between the Wolf Bay–Stull–Wunnumman Shear Zone in the Stull Lake–Edmund Lake greenstone belt and the North Kenyon Fault; and 3) Northern Superior superterrane (Skulski et al., 1999), the area north of the North Kenyon Fault. This report focuses on the Sachigo Lake to Yelling Lake corridor, including a relatively in-depth look at the Stull Lake–Edmund Lake greenstone belt and new geological data from a north-trending transect along the Stull River from Kistigan Lake to Curran Robinson Lake (Fig. GS-21-2).

MUNRO LAKE TERRANE

Thurston et al. (1991) defined the Munro Lake terrane as a crustal block bound by the Gods Lake Narrows Shear Zone in the north (equivalent in the map area to a southwestern splay of the Wolf Bay–Stull–Wunnumman Shear Zone), and the equivalent of the South Pipestone Lake Shear Zone and its eastern extensions in the south (Parmenter et al., 1999). The Gods Lake Narrows Shear Zone is described as a D_2 shear zone with dominant dextral transcurrent shear and a south-over-north component (Corkery et al., 1999). The shear zone deforms an earlier D_1 shear-zone fabric, with south-over-north thrust displacement, that deforms less than 2711 Ma conglomerates (Corkery et al., 1999; Corkery et al., GS-22, this volume).

At the southern edge of the transect area and the Munroe Lake terrane, the Sachigo Lake–Ponask Lake greenstone belt includes relatively highly strained rocks that form three south-facing assemblages (Fig. GS-21-1; Stone et al., 1996, 1997). In the north, at Ponask Lake, a rift-margin assemblage is interpreted to consist of tonalitic basement with overlying less than 2865 Ma tonalite- and komatiite-bearing conglomerate (D.W. Davis and M. Moore, Geochronology in the western Superior Province, unpublished report, Royal Ontario Museum, 1991), siltstone-sandstone (locally quartz arenite), minor amounts of calc-silicate marble and talcose-tremolitic komatiite. On western Ponask Lake, mafic and intermediate to felsic, tuffaceous to fragmental rocks and 2857 Ma felsic intrusions (D.W. Davis and M. Moore, Geochronology in the western Superior Province, unpublished report, Royal Ontario Museum, 1991) may represent an arc assemblage. Pillowed basalt occurs at Sachigo Lake and north of Ponask Lake.

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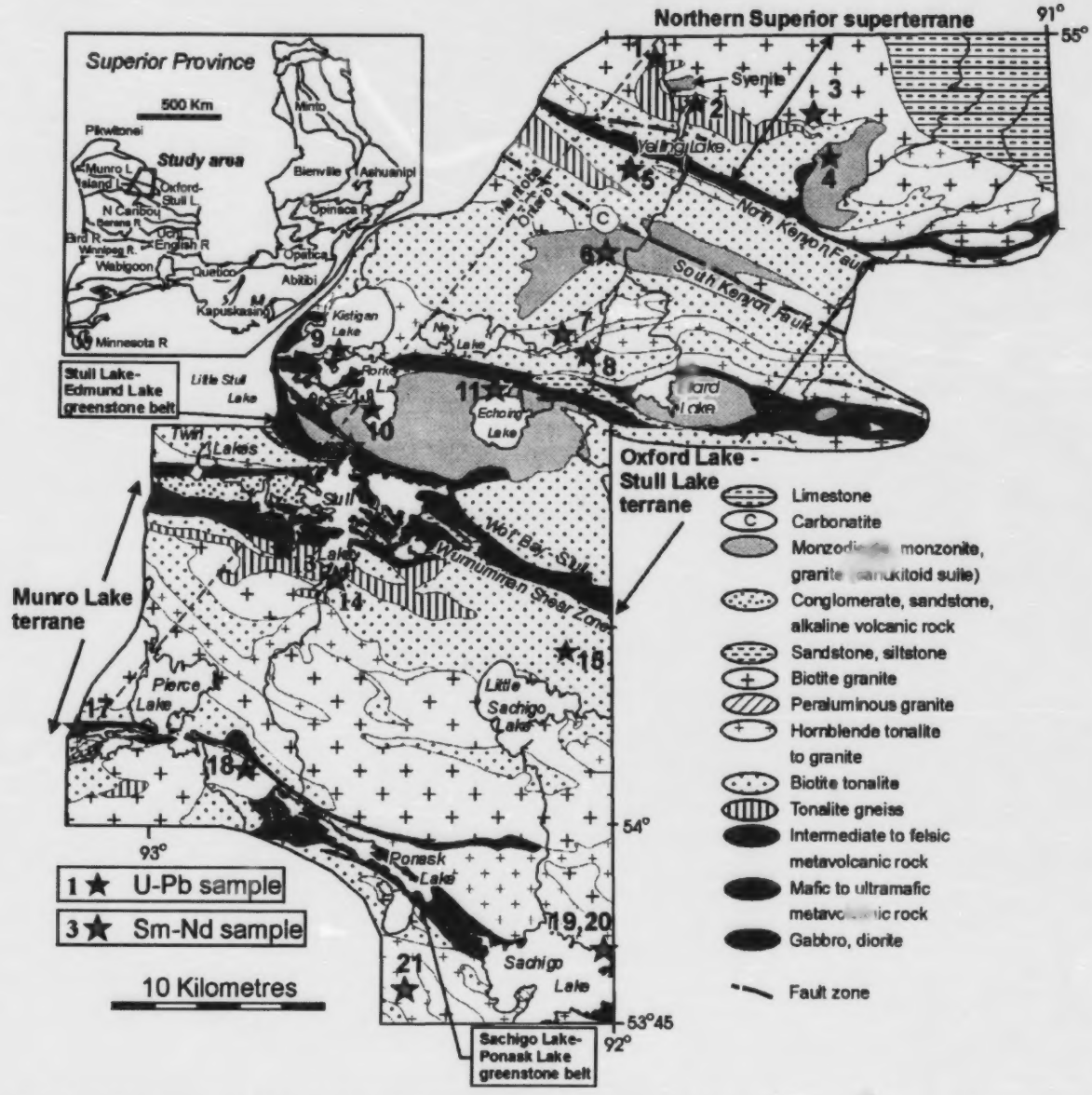

*Figure GS-21-1: Location map and geological map of the Sachigo Lake to Yelling Lake corridor. Inset map shows location within the northwestern Superior Province. Geological map shows principal terranes, geological units (*adapted from *Stone et al., 1999a), location of the Stull Lake–Edmund Lake greenstone belt, major shear zones and granitoid sample locations. Granitoid sample sites analyzed for both U-Pb and Sm-Nd for are shown as black stars; sites where only Sm-Nd isotopic data were collected are shown as grey stars.*

Hornblende-biotite tonalite gneiss occurs near the northern margin of the Munroe Lake terrane (Fig. GS-21-2). The contact between tonalite gneiss and basalt of the Hayes River Group (Stull Lake–Edmund Lake greenstone belt) in southern Stull Lake is obscured by glacial cover. The southernmost exposures of basalt are tectonically flattened, amphibolite-grade metabasalt. The tonalite gneiss is cut by multiple generations of syn- to post-tectonic granite pegmatite dykes and crosscut by boudinaged and tectonically dismembered mafic dykes. To the south, making up the central part of the Munroe Lake terrane, is a wide zone of moderately foliated biotite tonalite with enclaves of tonalite gneiss, amphibolite and hornblende tonalite. This tonalite is thought to be basement to the Sachigo–Ponask belt (Stone et al., 1996, 1997). The tonalite is intruded sequentially by intermediate hornblende tonalite to quartz diorite with lensoid inclusions of diorite and amphibolite, megacrystic hornblende granite, and massive to weakly foliated leucocratic biotite granite with inclusions of tonalite (Stone et al., 1996).

OXFORD LAKE–STULL LAKE TERRANE

This section focuses primarily on the Stull Lake–Edmund Lake greenstone belt and the internal and external granitoid plutons and major structures that define this terrane near the Ontario–Manitoba border.

Stull Lake–Edmund Lake Greenstone Belt

The Stull Lake–Edmund Lake greenstone belt comprises four tectonostratigraphic assemblages and four suites of internal plutonic rocks that record a complex history of ductile deformation and tectonic

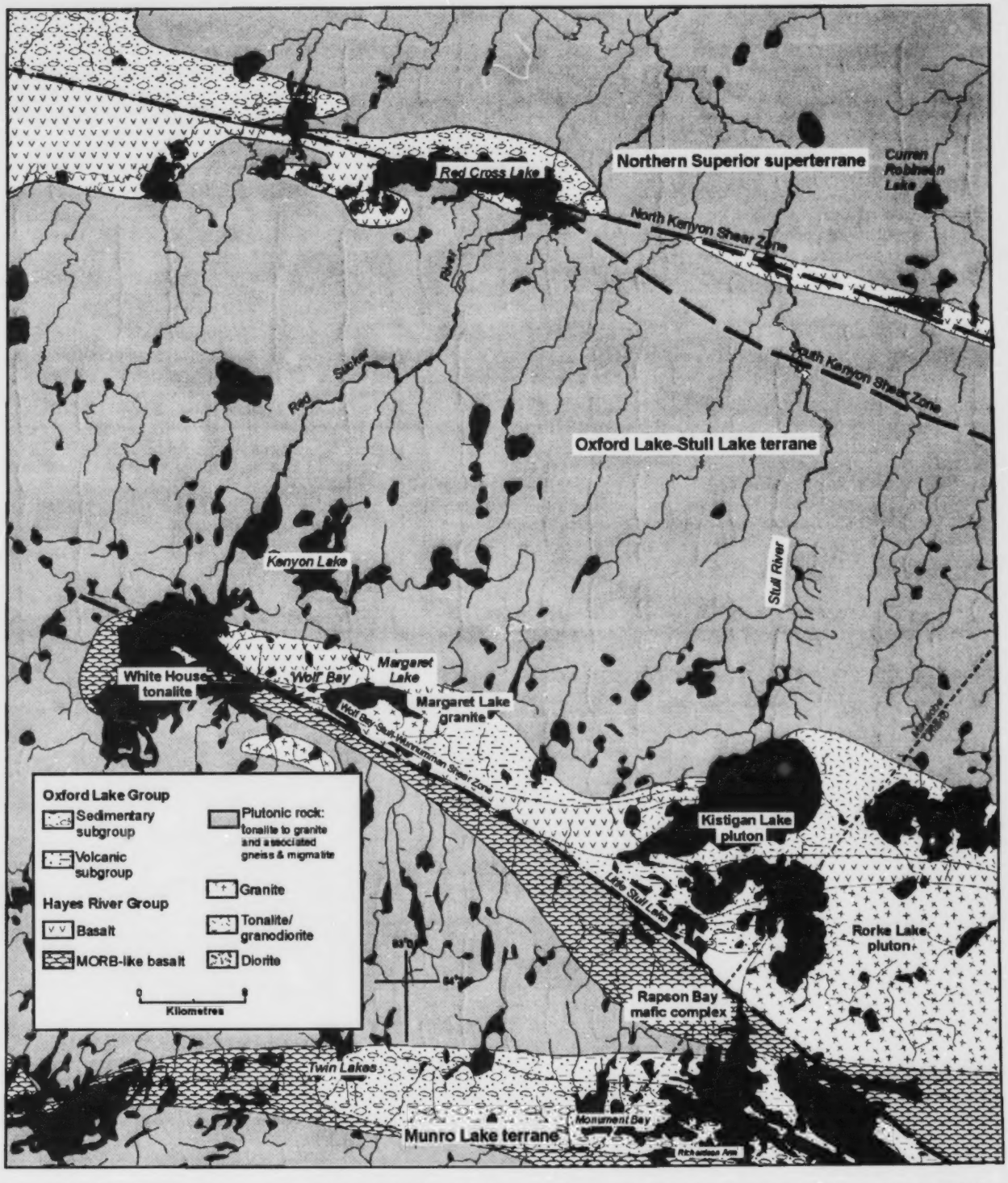

*Figure GS-21-2: Geological map of the Stull Lake–Edmund Lake greenstone belt, Oxford Lake–Stull Lake terrane (*adapted from *Corkery and Heaman, 1998).*

segmentation along the Wolf Bay–Stull–Wunnumman Shear Zone (WSWSZ; Fig. GS-21-2, -3). Available field, geochemical and geochronological constraints (Corkery and Heaman, 1998; Corkery and Skulski, 1998) allow us to subdivide the supracrustal rocks into a provisional tectonostratigraphy (pending ongoing geochronological studies at the Geological Survey of Canada [GSC] and Royal Ontario Museum [ROM]). These include:

1) south of the WSWSZ, greater than 2734 Ma Hayes River Group submarine tholeiitic basalt with depleted trace-element signatures;
2) north of the WSWSZ, greater than 2728 Ma Hayes River Group submarine tholeiitic basalt and andesite with flat light rare-earth element (LREE)–enriched, primitive-mantle–normalized patterns and positive Th/Nb values;
3) north of the WSWSZ, ca. 2741 to 2726 Ma Oxford Lake volcanic subgroup, calc-alkaline to shoshonitic, intermediate to felsic fragmental volcanic rocks and volcanic-derived sedimentary rocks; and
4) Oxford Lake sedimentary subgroup, fluvial-clastic sandstone and polymictic conglomerate.

Hayes River Group

Tholeiitic, sparsely vesicular, aphyric pillow basalt intruded by up to 50% gabbro sills forms the Rapson Bay mafic complex of the Hayes River Group, south of the WSWSZ (Fig. GS-21-3). This complex can be traced from Stull Lake in Ontario to the southern shore of Edmund Lake in Manitoba (Corkery and Skulski, 1998; Corkery and Heaman, 1998). The pillow basalt and gabbroic rocks are characterized by LREE- and large-ion lithophile element (LILE)–depleted, primitive-mantle–normalized profiles, similar to modern ocean-floor basalt (Corkery and Skulski, 1998). In the Edmund Lake area, a minimum age on the Rapson Bay mafic complex is provided by the crosscutting 2734 ± 2 Ma White House tonalite (Heaman and Corkery, 1998; Fig. GS-21-2).

Pillowed and massive tholeiitic basalt and rare andesite north of the WSWSZ are interbedded sporadically with oxide-facies banded iron-formation. The basalt includes LREE-enriched varieties with large negative Nb anomalies, as well as rare depleted tholeiitic basalt and basalt with flat primitive-mantle–normalized patterns (Corkery and Skulski, 1998). In the Edmund Lake area, LREE-enriched basalt has a minimum age defined by the crosscutting 2728 ± 2 Ma Margaret Lake granite (Fig. GS-21-2; Corkery and Heaman, 1998). The Hayes River Group volcanic rocks in the Stull Lake–Edmund Lake greenstone belt are chemically similar to diverse Hayes River Group basalt, found in the Knee Lake area, that is locally associated with 2.83 Ga rhyolite (Syme et al., 1998).

Oxford Lake volcanic subgroup

The White House tonalite and Margaret Lake granite are synvolcanic with Oxford Lake Group volcanism, and their age and field setting establishes that the Oxford Lake Group was built on a basement of Hayes River Group basalt (Corkery and Heaman, 1998). Four marine lithological associations are identified in the Oxford Lake volcanic subgroup on Little Stull Lake, including: Lodge Bay aphyric, intermediate to felsic tuff and breccia; Sickle Bay greywacke turbidite, argillite and oxide-facies iron-formation; Minnow Bay feldspar±quartz–phyric, intermediate to felsic tuff and breccia; and hornblende-phyric, high-K andesite, trachybasalt, trachyte and derived volcanogenic and epiclastic sedimentary rocks (Fig. GS-21-3; Corkery and Skulski, 1998; Stone et al., 1999a).

Corkery and Skulski (1998) reported a maximum age of 2726 ± 2 Ma (inheritance to 2751 ± 2 Ma) for a Lodge Bay aphyric rhyolite tuff interbedded with fine-grained, reworked, volcanogenic sedimentary rocks on Little Stull Lake. The Lodge Bay volcanic sequence is cut by a 2717 ± 1 Ma feldspar±quartz–phyric tonalite (Fig. GS-21-3; D.W. Davis and M. Moore, Geochronology in the western Superior Province, unpublished report, Royal Ontario Museum, 1991), thus bracketing deposition of aphyric volcanic rocks, argillite and iron-formation between 2726 Ma and 2717 Ma. The aphyric volcanic sequence is calc-alkaline and characterized by LREE- and LILE-enriched, and heavy rare-earth element (HREE)–depleted primitive-mantle–normalized profiles with

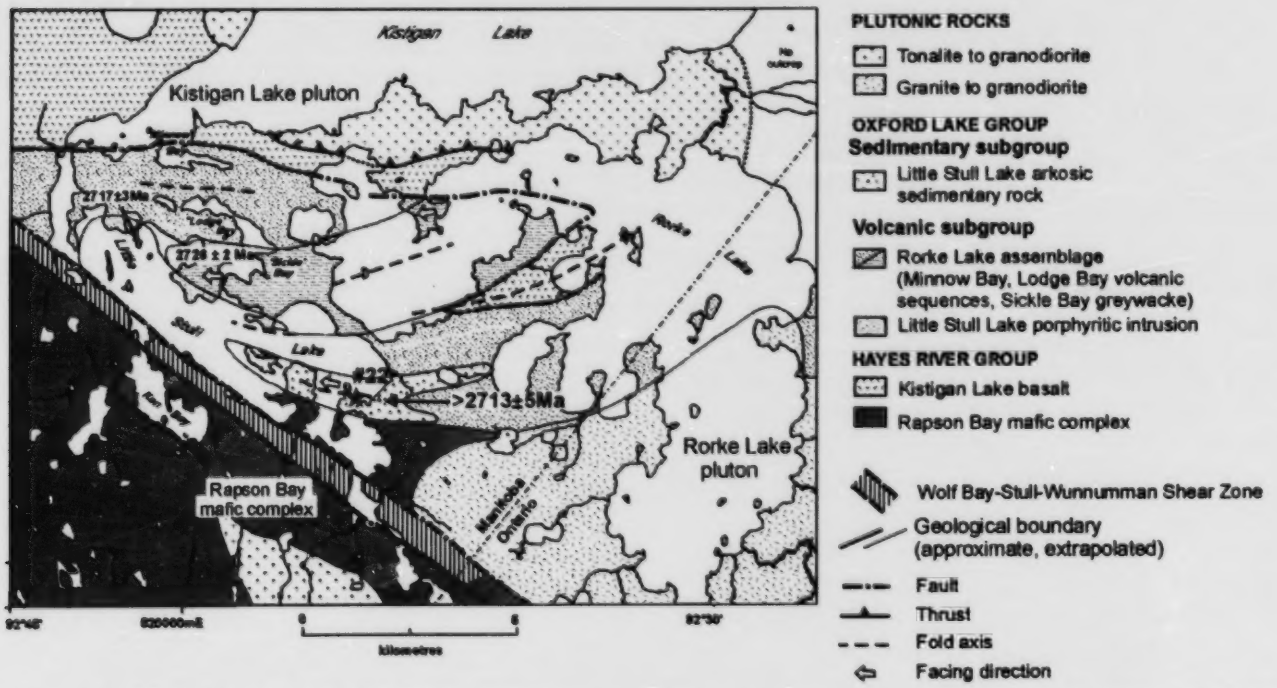

Figure GS-21-3: Geological map of the Stull Lake segment of the Stull Lake–Edmund Lake greenstone belt (adapted from *Corkery and Skulski, 1998 and Stone et al., 1999a*). *Black star indicates location of Oxford Lake sedimentary subgroup arkose sample (no. 22) for SHRIMP geochronology.*

negative Nb anomalies (Corkery and Skulski, 1998). Dacitic tuff, at Ellard Lake to the east (Fig. GS-21-1), has a U-Pb age of 2732 ± 1 Ma (Stone et al., 1999b) and is correlative with the Oxford Lake volcanic subgroup.

The Minnow Bay feldspar-phyric and quartz-feldspar–phyric andesite to rhyolite and associated sedimentary rocks are interpreted to be in fault contact with the Lodge Bay sequence (Corkery and Skulski, 1998). A minimum age on the Minnow Bay sequence is given by the crosscutting 2717 Ma feldspar±quartz–phyric tonalite (Fig. GS-21-3). The Minnow Bay sequence is calc-alkaline and shows trace-element signatures similar to those of the aphyric Lodge Bay sequence.

Intermediate to mafic fragmental rocks, which are locally hornblende-phyric on Rorke Lake (Fig. GS-21-3), include tuff, lapilli tuff and tuff breccia interlayered with reworked volcaniclastic and epiclastic rocks. Similar hornblende-phyric, mafic volcanic rocks and feldspar-phyric felsic volcanic rocks on Stull Lake, south of the WSWSZ, are mildly alkaline and include trachybasalt and trachyte (Fig. GS-21-3; Stone et al., 1999a). A sample of hornblende-phyric amygdaloidal trachybasalt from Stull Lake is characterized by pronounced enrichment in Th, LREE and LILE relative to other volcanic rocks in the Oxford Lake Group (Corkery and Skulski, 1998).

Oxford Lake sedimentary subgroup

Crossbedded and graded arkose, arkosic wacke and polymictic conglomerate constitute the Oxford Lake sedimentary subgroup and are exposed both north and south of the WSWSZ on Little Stull Lake and Stull Lake (Corkery and Skulski, 1998; Stone et al., 1999a). On Little Stull Lake, crossbedded arkose (foresets up to 5 cm) is overlain by polymictic, volcanic-derived conglomerate. On Stull Lake, polymictic conglomerate includes clasts of iron-formation, volcanic rocks, plutonic rocks and previously foliated granitoid and mylonitic rocks. This conglomerate overlies trachybasalt and trachyte (Stone et al., 1999a), and it is uncertain whether the interface between them marks an unconformity or, as suggested by Stone et al. (1999a), a conformable package of the Oxford Lake Group sedimentary subgroup. Large-scale (>2 cm), trough-type crossbedding, channel-facies conglomerate, and pebble lag deposits characterize the sedimentary subgroup and are characteristic of fluvial-clastic sedimentary rocks (Corkery and Skulski, 1998). The change in depositional setting from marine in the Oxford Lake volcanic subgroup (ca. 2726 Ma) to fluvial-subaerial in the sedimentary subgroup (<2713 Ma; *see* below), and the presence of polymictic conglomerate with clasts of underlying rock types in the latter, is evidence of an unconformable contact between the volcanic and sedimentary subgroups.

Internal Plutons

Four suites of internal plutons are recognized in the Stull Lake–Edmund Lake greenstone belt. These include 1) synvolcanic (Hayes River Group) gabbro and anorthositic gabbro plutons; 2) syn- to late volcanic (Oxford Lake volcanic subgroup), porphyritic tonalite, granodiorite and granite; 3) pre- to syntectonic (Oxford Lake volcanic subgroup) hornblende–biotite–K-feldspar porphyritic granodiorite; and 4) a sanukitoid suite.

Up to 50% of the Hayes River Group south of the WSWSZ on Little Stull Lake consists of synvolcanic gabbro sills that constitute the Rapson Bay mafic complex (Fig. GS-21-2; Corkery et al., 1997). These gabbroic rocks are associated with the tholeiitic basalt that extends from Rapson Bay in Ontario to southern Edmund Lake in Manitoba. The gabbro sills are medium grained, equigranular and locally plagioclase-phyric. An 800 m wide sill on Ken Bay in Little Stull Lake retains decimetre-scale layering of melagabbro, leucogabbro and anorthositic gabbro (Fig. GS-21-3).

Syn- to late volcanic plutons include the 2734 ± 2 Ma White House tonalite and the 2728 ± 2 Ma Margaret Lake granite (Corkery and Heaman, 1998) in the Edmund Lake area, and the 2717 ± 3 Ma feldspar±quartz–phyric tonalite on Little Stull Lake (Fig. GS-21-2; Corkery et al., 1997). The White House tonalite is a strongly foliated and recrystallized intrusion, dominated by fine-grained, equigranular biotite tonalite, that intrudes the Hayes River tholeiitic basalt described above (Corkery and Heaman, 1998). The tonalite is calc-alkaline and has primitive-mantle–normalized LREE enrichment and negative Nb anomalies similar to those found in felsic volcanic rocks of the Oxford Lake Group. The Margaret Lake granite is a fine- to medium-grained, equigranular, biotite leucogranite that is weakly foliated to nonfoliated (Corkery, 1996). The Little Stull Lake porphyritic intrusion is a tabular, 5 km long, composite intrusion of feldspar-phyric tonalite to quartz-feldspar–phyric trondhjemite. The pluton is moderately foliated, and the intensity of deformation increases toward the WSWSZ, which locally defines its western boundary.

The Kistigan Lake pluton is a large pre- to syntectonic hornblende–biotite–K-feldspar porphyritic granodiorite to hornblende-biotite tonalite that intrudes, and is in fault contact with (*see* below), the northern margin of the greenstone belt along the south shore of Kistigan Lake (Fig. GS-21-2, -3). This well foliated plutonic body is subdivided into three phases (Corkery et al., 1997): a predominant, coarse-grained, equigranular tonalite to granodiorite phase; a finer grained, mafic tonalite to quartz diorite marginal phase; and a local, K-feldspar megacrystic (1–7 cm), medium- to coarse-grained granodiorite.

The Rorke Lake pluton is a relatively massive, K-feldspar megacrystic granodiorite and hornblende-biotite granodiorite that intrudes the greenstone belt in the Rorke Lake area (Fig. GS-21-2, -3). This pluton is part of a monzodiorite-monzonite-granite sanukitoid suite (Stone et al., 1997). The sanukitoid bodies are typically late, oval-shaped, zoned plutons that have mafic marginal zones (diorite, monzodiorite and monzonite) containing crustal xenoliths, grading into more differentiated granodiorite and granite in the core. The geochemistry of this suite is characterized by high Mg#s, high Cr and Ni concentrations, and enrichment in LREE and LILE relative to other granitoid rocks (cf. Stern and Hanson, 1991).

Structural History of the Stull Lake–Edmund Lake Greenstone Belt

The Stull Lake–Edmund Lake greenstone belt is cut by the WSWSZ, a large northeast-trending deformation zone that divides the belt into two domains: a southern domain dominated by depleted tholeiitic mafic volcanic rocks of the Hayes River Group, and a northern domain comprising tholeiitic Hayes River volcanic rocks and fault-bounded panels of the younger Oxford Lake volcanic subgroup (Fig. GS-21-2; Corkery et al., 1997; Corkery and Skulski, 1998; Jiang and Corkery, 1998). The Oxford Lake sedimentary subgroup is an element common to both sides of the deformation zone. Jiang and Corkery (1998) recently summarized the structural evolution of the greenstone belt and recognized four sets of ductile structures.

Structures belonging to D_1 are confined to tholeiitic gabbro and basalt of the Hayes River Group on the south shore of Edmund Lake (Fig. GS-21-3; Jiang and Corkery, 1998). These structures comprise a shallow-dipping S_1 foliation and shallow-plunging L_1 stretching lineation in deformed gabbro and basalt. The S_1 foliation is a shape fabric in deformed gabbro and a compositional, tectonic layering of mylonitized gabbro and basalt that represents the C-foliation in ductile shear zones. The sense of shear in these early deformation zones is not evident. These early structures are overprinted by the D_{2-4} structures described below.

Structures belonging to D_2 reflect regional isoclinal folding, with map-scale F_2 folds in the Little Stull Lake area that primarily control the map pattern (Fig. GS-21-3; Jiang and Corkery, 1998). The F_2 folds have gentle plunges, are slightly overturned to the southeast, and have

east-northeast trends that curve into parallelism close to the southeast-trending D_3 WSWSZ. Map-scale F_2 structures include an anticline in the Oxford Lake sedimentary subgroup, an anticline in the Minnow Bay sequence, and a syncline and anticline in the Sickle Bay–Lodge Bay sequences. Jiang and Corkery (1998) described S_2 as a differentiated crenulation cleavage that overprints S_1 cleavage in the Rapson Bay mafic complex.

The D_3 deformation reflects noncoaxial dextral transpression along the regional WSWSZ (Fig. GS-21-2). Jiang and Corkery (1998) suggested that it is a continuation of D_2, reflecting an evolution from folding to dextral transpression. This large structure, and kinematically similar small-scale shear zones in the Rapson Bay mafic complex, strike 110 to 130° and have S-C structures and other kinematic indicators that indicate a dextral sense of shear (Corkery and Skulski, 1997; Jiang and Corkery, 1998). Lineations within the WSWSZ and parallel allied structures are subhorizontal, whereas they are steeper and more variable in less deformed wall rocks.

Structures belonging to D_4 reflect north-south compression. They are characterized by 1) small-scale, cleavageless Z-folds of D_3 mylonite in the WSWSZ, with geometries that are inconsistent with dextral transpression, and 2) S-folds along southern Kistigan Lake, which cannot be attributed to parasitic F_2 or D_3 drag folds (Jiang and Corkery, 1998). On Kistigan Lake, a shallow north-dipping D_4 shear zone, with a mylonitic foliation and a north-plunging L_4 stretching lineation, separates the Kistigan Lake pluton to the north from Hayes River Group basalt, suggesting north-side-up thrusting.

Granitoid Plutons North of the Stull Lake–Edmund Lake Greenstone Belt

Three lithological suites of rocks dominate the plutonic domain that extends from north of the Stull Lake–Edmund Lake greenstone belt to the North Kenyon Fault (Fig. GS-21-1). An early complex orthogneiss, generally of tonalitic composition, is intruded by a younger, foliated, biotite tonalite-granodiorite that is, in turn, intruded by biotite granite, hornblende-bearing granite and associated late aplitic and pegmatitic dykes.

The early gneiss complex occurs both as the dominant rock type over kilometre-scale sections of the Stull River transect and as xenoliths in younger intrusive rocks. It generally occurs as a centimetre- to decametre-scale layered biotite or rarely hornblende tonalite gneiss. Fine- to medium-grained grey quartzofeldspathic bands form original restite, with 5 to 20% leucocratic mobilizate. A *lit* injection of light grey biotite tonalite forms the youngest phase of the early gneiss. In several locations, the dominant mafic mineral is hornblende or a retrograded mix of biotite, chlorite and hornblende remnants. The distribution of mafic minerals as clots in many locations is interpreted to indicate a younger period of metamorphism that postdates the early migmatite development.

Granodiorite to biotite tonalite is the dominant rock type along both the Stull River transect and the 1:50 000 scale corridor near the provincial border. These plutonic massifs vary from coarse- to medium-grained biotite granodiorite, locally microcline megacrystic, to medium- to fine-grained tonalite. This unit intrudes and truncates the gneissic layering and isoclinal folding in the early gneiss complex. However, there is local preservation of hornblende and an indication of late retrograde metamorphism similar to that observed in the gneiss.

Medium- to coarse-grained biotite granite forms from a few per cent up to (rarely) 50% of outcrops. Fine-grained aplitic and pegmatitic dykes are also common.

The early gneiss is generally well foliated to strongly lineated and was migmatitic prior to tonalite injection. However, both of these units share a strong east-west fabric throughout most of the Stull River transect, indicating at least two metamorphic events. The younger granite suite generally occurs as straight-walled intrusions that truncate the east-west fabric with a weak foliation.

In the area of the South and North Kenyon faults (Fig. GS-21-1), all units are strongly sheared and a fabric trending 300 to 310° transposes or overprints all earlier fabrics. The South Kenyon Fault is believed to be a splay of the North Kenyon Fault, and is a north-northwest-trending, 480 km long, subvertical deformation zone characterized by broad, greenschist facies, 1 to 2 km wide mylonite zones (Osmani and Stott, 1988; Stone et al., 1999b).

NORTH KENYON FAULT AND THE NORTHERN SUPERIOR SUPERTERRANE

The region north of the North Kenyon Fault comprises five plutonic suites (Stone et al., 1999b). A belt-like domain of medium-grained biotite tonalite gneiss, locally containing plagioclase porphyroclasts, is exposed north of Yelling Lake (Fig. GS-21-1). Well foliated biotite monzogranite is the most abundant rock type in this area and occurs as strongly foliated, K-feldspar megacrystic syenogranite on the north shore of Curran Robinson Lake in the west (Fig. GS-21-2). A sample of sheared granodiorite within the North Kenyon Fault at Yelling Lake is dated at 2722 ± 2 Ma (Fig. GS-21-1; Stone et al., 1999b), providing a maximum age on shear-zone deformation. Large tabular bodies of hornblende tonalite curve into parallelism with the North Kenyon Fault as the boundary of the fault is approached, reflecting an apparent dextral sense of drag. An oval-shaped pluton of well foliated hornblende–biotite–K-feldspar megacrystic granodiorite, belonging to the sanukitoid suite, intrudes biotite tonalite and biotite granite east of Yelling Lake (Fig. GS-21-1).

The North Kenyon Fault on Yelling Lake comprises a 1 km wide protomylonite to mylonite zone within granodiorite (Osmani and Stott, 1988). The rock matrix is reported to be extensively altered to chlorite and epidote, and feldspar augen are saussuritized. Asymmetrical quartz aggregates with subhorizontal long axes in the granodiorite suggest dextral noncoaxial flow (Osmani and Stott, 1988). Planar fabric within the mylonite dips northerly, and stretching lineations are mainly subhorizontal, although steep down-dip lineations are locally present (Stone et al., 1999a).

GEOCHRONOLOGICAL AND ND ISOTOPIC RESULTS

New geochronological and Nd isotopic data are summarized here from both the Stull Lake–Edmund Lake greenstone belt and a regional survey of granitoid rocks extending from south of the WSWSZ, near Sachigo Lake, to near the North Kenyon Fault in the Yelling Lake area. Zircons from sedimentary and plutonic rocks were analyzed with the GSC's SHRIMP II facility, using extraction and analytical methods described in Stern (1997). The Sm-Nd isotopic data on volcanic and plutonic rocks were collected in the geochronological laboratories of the GSC using methods described in Skulski et al. (1996).

Nd Isotopic Composition of Granitoid Rocks

Figure GS-21-4 shows Nd model ages (DePaolo [1981] depleted-mantle model) of granitoid rocks along the Sachigo Lake to Yelling Lake corridor. Differences in Nd model ages are apparent between the three crustal domains, indicating distinct crustal histories. In the Munro Lake terrane, Nd model ages range from 3.02 Ga in tonalite gneiss south of the Stull Lake belt to 2.92 Ga in hornblende-biotite tonalite north of the Sachigo Lake–Ponask Lake greenstone belt. In contrast, Nd model ages in the Oxford Lake–Stull Lake terrane are largely Neoarchean and range from 2.80 Ga in hornblende-biotite granodiorite of the sanukitoid suite to 2.77 Ga in moderately foliated hornblende-biotite granodiorite that flanks the Stull Lake–Edmund Lake greenstone belt. Older Nd model ages of 2.94 Ga are found in biotite tonalite in the northern parts of the

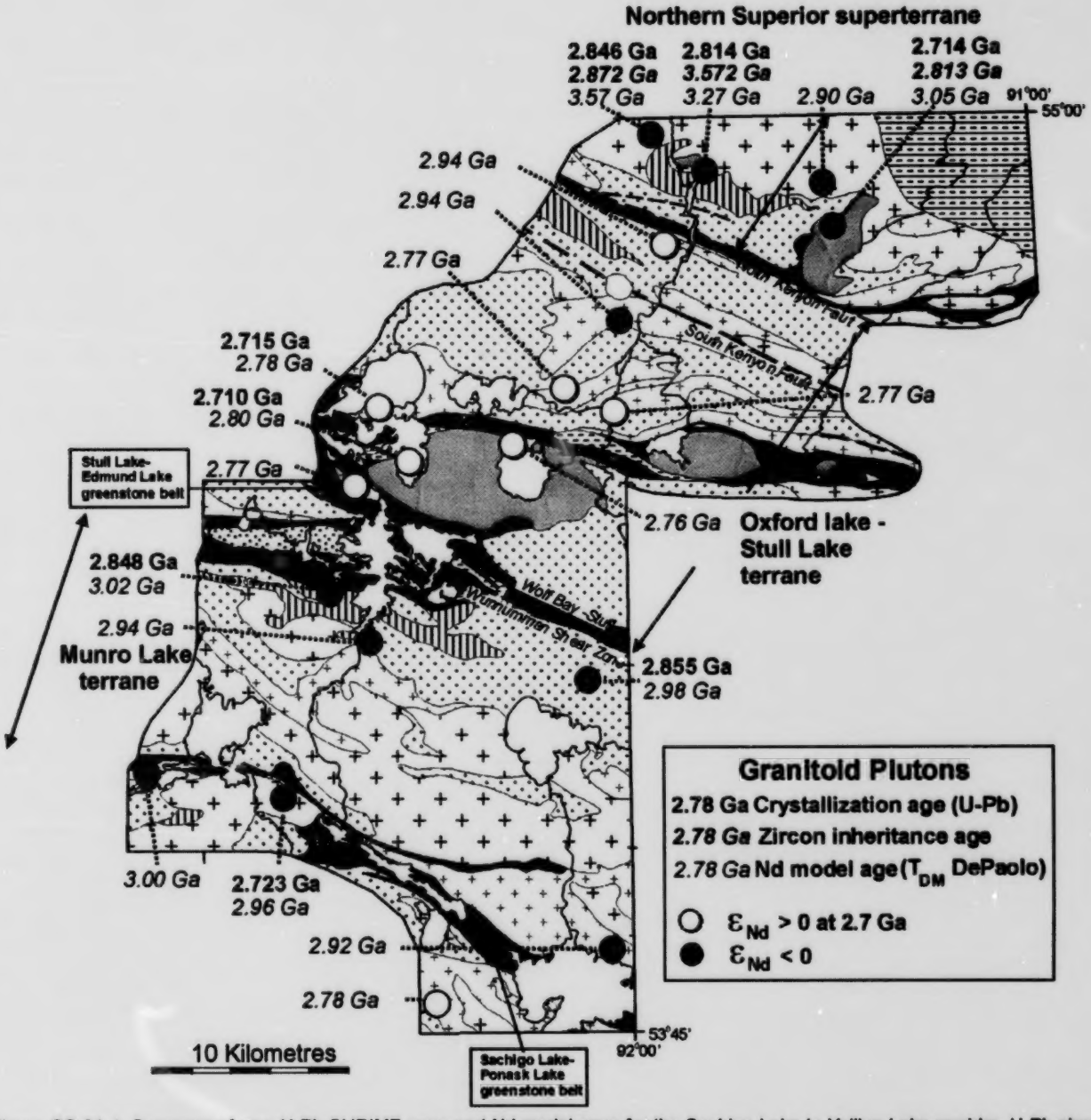

Figure GS-21-4: Summary of new U-Pb SHRIMP ages and Nd model ages for the Sachigo Lake to Yelling Lake corridor. U-Pb zircon SHRIMP ages are shown in bold and include crystallization ages (normal type) and maximum age of inheritance (bold italic type). Nd model ages (DePaolo [1981] depleted-mantle model) of whole-rock plutonic samples are shown in italic type. Sample locations are shown on Figure GS-21-1; rock types for dated samples are given in Figures GS-21-6 to -8.

terrane. Paleoarchean to Mesoarchean Nd model ages are found across the North Kenyon Fault in the Northern Superior superterrane. These range from 3.57 to 2.90 Ga in foliated biotite granite, to 3.27 Ga in tonalite gneiss, and 3.05 Ga in foliated hornblende-biotite granodiorite of the sanukitoid suite.

Nd Isotopic Composition of Volcanic Rocks

Eleven whole-rock samples of volcanic rocks were analyzed for their Nd isotopic composition in the Stull Lake–Edmund Lake greenstone belt (Fig. GS-21-5). Three tholeiitic basalt samples from the Hayes River Group south of the WSWSZ are depleted in incompatible LREE and have relatively high initial ε_{Nd} values, ranging from +3.8 to +0.5, that reflect derivation from a depleted-mantle source (generally accepted to be greater than +2.0 at ca. 2.7 Ga). These results are consistent with an ocean-floor or thin continental margin origin for this submarine mafic volcanic sequence.

Three samples of Hayes River Group basalt from north of the WSWSZ, characterized by LREE enrichment and negative Nb anomalies, have ε_{Nd} values ranging from +2.3 to -0.6. These lower ε_{Nd} values may reflect recycling of older continental crust in a subduction-zone setting.

A sample of 2726 Ma aphyric dacite from the Lodge Bay sequence of the Oxford Lake volcanic subgroup has an ε_{Nd} value of +1.4. The

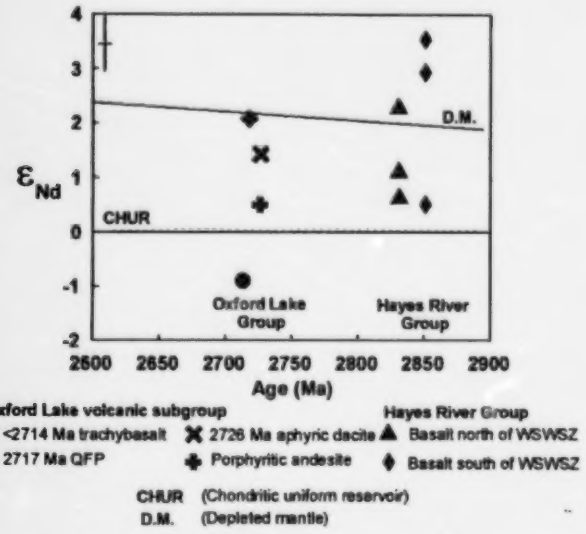

Figure GS-21-5: Epsilon Nd (ε_{Nd}) value versus age for whole-rock volcanic samples from the Stull Lake–Edmund Lake greenstone belt.

slightly younger 2717 Ma porphyritic tonalite intrusion on Little Stull Lake also has a positive ε_{Nd} value of +2. These data suggest that calc-alkaline, LREE-enriched magmas with negative Nb anomalies did not interact significantly with older continental crust. This observation is consistent with the field observation that synvolcanic intrusions (2734–2728 Ma) on Edmund Lake cut relatively juvenile mafic crust of the Hayes River Group.

Hornblende-phyric high-K andesite north of the WSWSZ, and trachybasalt underlying conglomerate of the Oxford Lake sedimentary subgroup south of the WSWSZ, are characterized by low ε_{Nd} values of +0.5 and -0.9, respectively. These low ε_{Nd} values may reflect a fundamental change in the source of high-K magmatism that involves the recycling of older continental crust.

In general, pre-2717 Ma volcanic rocks of the Stull Lake–Edmund Lake greenstone belt have relatively juvenile Nd isotopic compositions, as do the internal granitoid plutons that cut the belt. These data suggest that the volcanic belt developed on either a thin continental margin or in an oceanic setting. A major change in the Nd isotopic composition of mildly alkaline and high-K volcanic rocks occurs just prior to the deposition of synorogenic conglomerate of the Oxford Lake sedimentary subgroup. This change is recorded in the first appearance of Mesoarchean and Paleoarchean detrital zircons in late sedimentary rocks of the Stull Lake–Edmund Lake greenstone belt (below) and from correlative units in the Oxford Lake–Stull Lake terrane to the west.

U-Pb Geochronology

Eight samples of granitoid rocks were dated, using the SHRIMP, to investigate the crystallization ages of, and possible zircon-inheritance patterns from, widely spaced representative units with contrasting Nd model ages (Fig. GS-21-4). In addition, detrital zircons from a sample of crossbedded arkose from the Oxford Lake sedimentary subgroup were dated with the SHRIMP as well.

Munro Lake terrane

Three rocks were dated in the Munro Lake terrane, two samples of tonalite gneiss and a moderately foliated K-feldspar–biotite granodiorite (Fig. GS-21-4). Ten zircon grains from a biotite tonalite gneiss from south of the WSWSZ (sample 15; UTM 15, 560500E, 5907700N) have a U-Pb age of 2855 ± 5 Ma (Fig. GS-21-6a). This rock has a Nd model age of 2.98 Ga (initial ε_{Nd} of +0.15), although it shows no evidence of inherited zircons of this age. A tonalite gneiss from south of the Stull Lake–Edmund Lake greenstone belt (sample 13; UTM 15, 519752E, 6022590N) has an age of 2848 ± 7 Ma (10 zircons analyzed) and a 3.02 Ga Nd model age (initial ε_{Nd} of -0.28; Fig. GS-21-6b). The Nd data suggest that ca. 3 Ga crust may have been implicated in the source of tonalitic magmas. Data from other parts of the Munroe Lake terrane show similar results. To the west, in the Munro Lake terrane south of the Knee Lake greenstone belt, a felsic gneiss has an age of 2883 Ma (Corkery et al., 1999). Synvolcanic quartz porphyry, which cuts mafic rocks in the Ponask Lake belt, is dated at 2857 ± 2 Ma (D.W. Davis and M. Moore, Geochronology in the western Superior Province, unpublished report, Royal Ontario Museum, 1991) and is contemporaneous with tonalite gneiss to the north. Early quartz-rich conglomerate in the Ponask Lake belt that rests unconformably on tonalite (Stone et al., 1996) has a unimodal detrital zircon age of 2865 ± 1 Ma (D.W. Davis and M. Moore, Geochronology in the western Superior Province, unpublished report, Royal Ontario Museum, 1991), 10 m.y. older than the biotite tonalite gneiss dated here.

A K-feldspar–biotite granodiorite (sample 18; UTM 15, 513200E, 5992150N), which intrudes the Ponask Lake–Sachigo Lake greenstone belt, has a U-Pb age of 2721 ± 6 Ma (Fig. GS-21-1, -6c). This rock has a 2.96 Ga Nd model age (initial ε_{Nd} value of -0.70), which may reflect an older ca. 3 Ga crustal source beneath the Munro Lake terrane.

Oxford Lake–Stull Lake terrane

Uranium-lead ages were obtained by SHRIMP for the Kistigan Lake and Rorke Lake plutons and for a sample of crossbedded arkose from the Oxford Lake sedimentary subgroup in the Stull Lake–Edmund Lake greenstone belt (Fig. GS-21-3, -4). Eleven zircons from a hornblende–biotite–K-feldspar granodiorite phase of the Kistigan Lake pluton (sample 9; UTM 15, 524789E, 6051127N) give a U-Pb age of 2715 ± 8 Ma (Fig. GS-21-1, -7a). A single zircon core had a $^{207}Pb/^{206}Pb$ age of 2747 Ma that, along with the 2.78 Ga Nd model age (initial ε_{Nd} of +1.61), indicates a juvenile source for the Kistigan Lake pluton. The age of this pluton also provides a maximum age estimate for D_4 thrusting of the Kistigan Lake pluton onto Hayes River basalt.

Ten zircons from a hornblende-biotite granodiorite phase of the Rorke Lake sanukitoid (sample 10; UTM 15, 531286E, 6043247N) give a U-Pb age of 2710 ± 6 Ma (Fig. GS-21-1, -7b). The absence of inherited zircons in this rock and a Nd model age of 2.77 Ga (initial ε_{Nd} of +1.15) lend further support to a juvenile source for this pluton. From crosscutting relationships, the age of the Rorke Lake pluton provides a minimum age for the Minnow Bay volcanic sequence, and a maximum age for the WSWSZ that cuts its southern margin.

Twenty-two detrital zircons from a sample of crossbedded arkose in the Oxford Lake sedimentary subgroup on Little Stull Lake (sample 22; UTM 15, 523010E, 6044495N) were analyzed with the SHRIMP (Fig. GS-21-3, -7c). Zircons in this sample range in age from 2905 ± 8 Ma to 2713 ± 5 Ma, the latter being a U-Pb age of five spots on a single zircon that constrains the maximum age of sedimentation. The sample also contains detrital zircons of 2842 ± 9 Ma, similar in age to tonalite gneiss in the Munro Lake terrane and Hayes River felsic volcanism at Knee Lake (ca. 2830 Ma; Corkery et al., GS-22, this volume), and younger detrital zircons of 2740 to 2720 Ma that may reflect derivation from sources in the underlying Oxford Lake volcanic subgroup. Detrital zircons in the fluvial-clastic facies of the Oxford Lake sedimentary subgroup in the Knee Lake greenstone belt range in age from 3647 Ma to 2711 ± 2 Ma on Gods Lake, and from 2798 Ma to 2707 +9/-8 Ma on southern Knee Lake (Corkery et al., GS-22, this volume), and thus provide similar maximum age constraints on sediment deposition. From the detrital zircon data, a maximum age for D_2 folding of the Oxford Lake sedimentary subgroup on Little Stull Lake is 2713 ± 5 Ma.

A number of lines of evidence are consistent with deposition of the

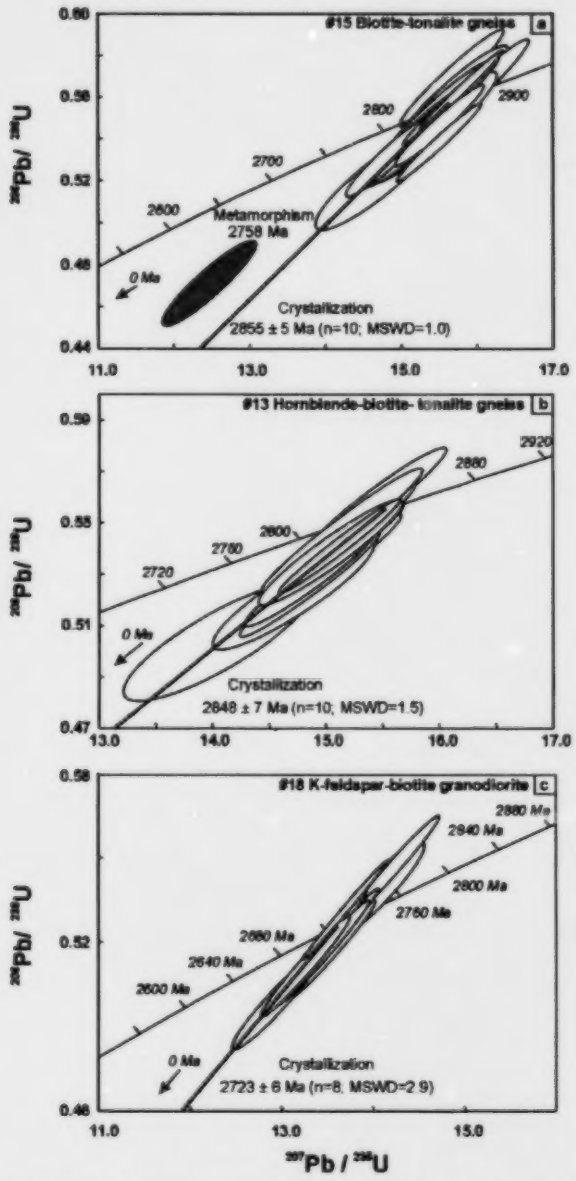

Figure GS-21-6: U-Pb concordia diagram of SHRIMP zircon data from plutonic rocks in the Munro Lake terrane. Sample locations are shown on Figure GS-21-1 (UTM coordinates in text). In all three diagrams, the crystallization age is given for regression through spot analyses (number of spots = n) that is forced through a zero-age intercept (MSWD is mean standard weighted deviation of regression). Filled (grey) ellipse in (a) is a single $^{207}Pb/^{206}Pb$ age determination on a metamorphic overgrowth in biotite tonalite gneiss.

sedimentary subgroup after emplacement of the 2710 ± 6 Ma Rorke Lake pluton. Mesoarchean detrital zircons in arkose (ca. 2.9 Ga) and low ε_{Nd} values in trachybasalt (-0.9) that underlies the conglomerate both indicate proximity of older continental crust, at less than 2713 ±5 Ma. This crust was not present when the 2710 ± 6 Ma Rorke Lake pluton (ε_{Nd} = +1.1) and earlier magmas were emplaced in the belt. Underthrusting of Mesoarchean crust beneath the Stull Lake belt after 2710 Ma could account for the contaminated nature of the high-K volcanic rocks, the change from marine to fluvial depositional setting, and the presence of previously deformed plutonic clasts in conglomerate of the Oxford Lake sedimentary subgroup.

Northern Superior superterrane

Three plutonic rocks from the Northern Superior superterrane that were dated using the SHRIMP show evidence of zircon inheritance and younger metamorphic events (Fig. GS-21-1, -4).

A sample of biotite monzogranite from northwest of Yelling Lake (sample 1; UTM 15, 569300E, 6090850N) has a complex geological history preserved in the zircon. Fifteen analyses of optically zoned interiors of zircon grains give a U-Pb crystallization age of 2846 ± 5 Ma (Fig. GS-21-1, -8a). A single, possible inherited core has a $^{207}Pb/^{206}Pb$ age of 2872 Ma. This rock has a Nd model age of 3.57 Ga (initial ε_{Nd} value of -6.2) and must have been derived by partial melting of Paleoarchean crust. Although Paleoarchean zircons were not observed in this sample, the tonalite gneiss to the south (*see* next paragraph) does contain inherited components of Paleoarchean age. Some of the grains have distinctive, thin, discontinuous, and unzoned overgrowths that have uniformly low Th/U values of approximately 0.038, characteristic of metamorphic zircon. Three of these overgrowths have $^{207}Pb/^{206}Pb$ ages in the range 2760 to 2717 Ma and give a minimum age of metamorphism of ca. 2717 Ma.

A biotite tonalite gneiss at the northern tip of Yelling Lake (sample 2; UTM 15, 576200E, 6085300N), which is cut by 2722 Ma granodiorite to the south, contains a complex history of zircon inheritance and metamorphic overgrowths for which SHRIMP dating is perfectly suited (Fig. GS-21-1, -8b). Zircons from this rock contain an abundance of inherited cores mantled by thick, optically zoned overgrowths and, in some cases, thin, unzoned, discontinuous rims. Nine analyses of zoned mantles give a U-Pb age of 2814 ± 4 Ma, interpreted to be the crystallization age of the igneous protolith. The $^{207}Pb/^{206}Pb$ ages of inherited cores range from 3572 to 3209 Ma. Two analyses of thin, unzoned rims give $^{207}Pb/^{206}Pb$ ages of 2744 and 2741 Ma. These rims are interpreted to represent metamorphic zircon growth, with a minimum age of 2741 Ma. The tonalite gneiss has a Nd model age of 3.27 Ga (initial ε_{Nd} value of -5.30). Both the direct evidence for zircon inheritance and the Nd isotopic data clearly indicate an ancient crustal source of early Mesoarchean to Paleoarchean age for this gneiss.

There is no evidence that Paleoarchean crust was implicated in either the magmatic source region or the sediment source in the adjacent Oxford Lake–Stull Lake terrane or terranes to the south, prior to deposition of the Oxford Lake sedimentary subgroup after ca. 2.71 Ga. At Gods Lake to the southwest, 3.6 Ga detrital zircons are found in less than 2711 Ma Oxford Lake sedimentary rocks (Corkery et al., GS-22, this volume) and 3.5 Ga detrital zircons are found in the less than 2709 Ma Cross Lake Group to the west (Corkery et al., 1992). The Northern Superior superterrane (Skulski et al., 1999) is interpreted to include the crust north of the North Kenyon Fault and to extend northwestward to the Assean Lake Block, where 3.54 Ga tonalite gneiss occurs within a crustal domain containing widespread Nd model ages of greater than 3.5 Ga (Böhm et al., 2000).

Zircons from a sample of a hornblende–biotite–K-feldspar granodiorite sanukitoid pluton east of Yelling Lake (sample 4; UTM 15, 595700E, 6077700N) were dated with the SHRIMP (Fig. GS-21-1, -8c). Eleven analyses of zoned grains, which either lack inherited cores or represent zoned mantles on cores, give a U-Pb age of 2714 ± 8 Ma, interpreted as the crystallization age of the pluton. Inherited zircons in this rock have $^{207}Pb/^{206}Pb$ ages ranging from 2813 to 2720 Ma. The Nd model age of this pluton is 3.06 Ga (initial ε_{Nd} value of -1.71) and, together with the zircon data, reflects recycling of Mesoarchean or older crust in its magmatic source region.

The absence of metamorphic zircon overgrowths in the sanukitoid pluton is interpreted to indicate that metamorphism in the Northern

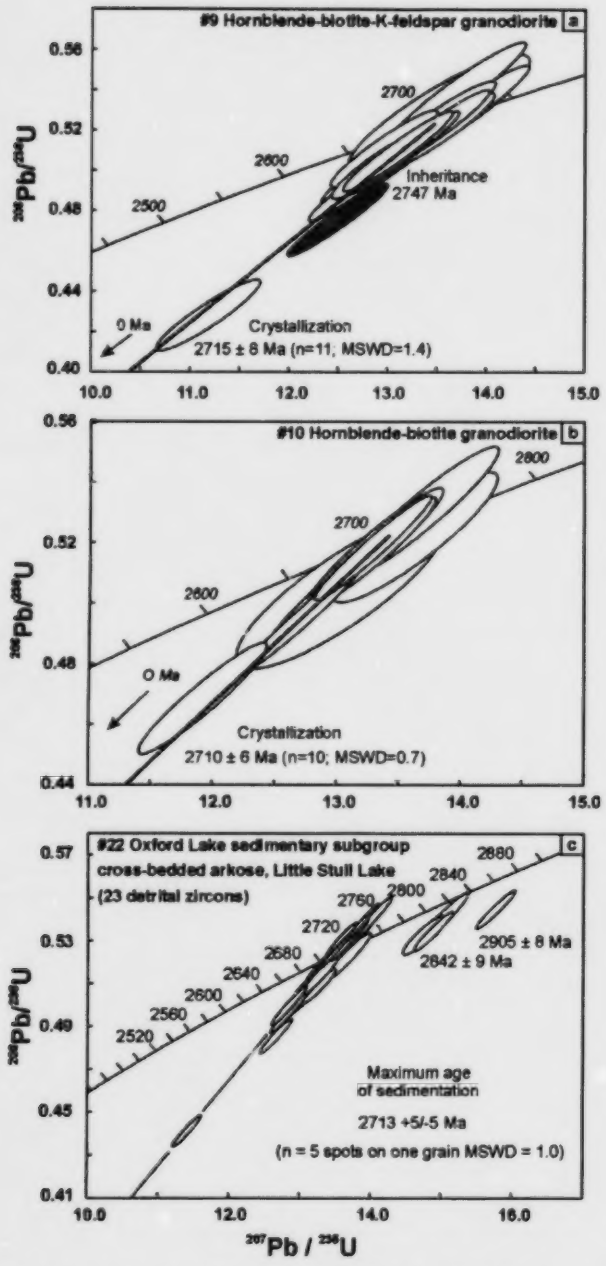

Figure GS-21-7: U-Pb concordia diagram of SHRIMP U-Pb zircon data from plutonic and sedimentary rocks in the Oxford Lake–Stull Lake terrane. Sample locations are shown on Figures GS-21-1 and -3 (UTM coordinates in text). In (a) and (b), the crystallization age is given for a regression forced through the origin, and the number of analyses and MSWD are given in parentheses. Filled (black) ellipse in (a) is a single $^{207}Pb/^{206}Pb$ age determination on an inherited zircon core. Twenty-three detrital zircon analyses are shown in (c). Five spots on the youngest grain were regressed (free regression) to yield a maximum age of sedimentation.

Superior superterrane can be bracketed between ca. 2.74 Ga (age of overgrowths in tonalite gneiss) and 2.71 Ga (age of sanukitoid). As mentioned earlier, the North Kenyon Fault cuts a 2722 Ma granodiorite at Yelling Lake, and movement along this fault could have been related closely in time with regional metamorphism of the Northern Superior superterrane.

TECTONIC SYNTHESIS

The data presented here provide important constraints for the tectonic evolution of the northwestern Superior Province.

Quartzite, marble and komatiite in the Sachigo Lake–Ponask Lake greenstone belt, although undated, were likely deposited after 2.865 Ga, the age of detrital zircons in basal conglomerate at Ponask Lake. These sedimentary rocks are interpreted to represent the vestiges of a northern platform sequence on the ca. 3 Ga North Caribou protocraton. Evidence of this 3 Ga protocraton is found in the older Nd model ages of plutonic rocks in the Munro Lake terrane.

Depleted submarine basalt of the Hayes River Group either represents slivers of exotic ocean-floor crust or continental margin obducted during D_1 deformation at Edmund Lake, or juvenile continental crust telescoped prior to 2734 Ma, the age of the crosscutting tonalite. The presence of both depleted basalt and possible contaminated arc basalt in the Hayes River Group on Kistigan lake may be an indication that these different volcanic rocks formed on the thin margin of the North Caribou protocontinent. A pre–Oxford Lake volcanic subgroup angular unconformity at Knee Lake (Syme et al., 1998), and occurrence of D_1 fabrics that is restricted to Hayes River basalt at Edmund Lake, may be evidence that this ca. 2.83 Ga continental-margin crust was tectonically shortened before Oxford Lake time (>2.73 Ga). Deposition of Oxford Lake volcanic and sedimentary rocks in a marine setting suggests that the earlier tectonic shortening was an accretionary event, and not the closure of an ocean basin.

Oxford Lake volcanism, at 2.73 to 2.72 Ga, is widespread within the Oxford Lake–Stull Lake terrane and extends as far west as Cross Lake. This calc-alkaline volcanic episode is interpreted to reflect the development of a continental-margin arc prior to collision of the Northern Superior superterrane. The transition from juvenile magmatic sources at 2.71 Ga, when the Rorke Lake pluton was emplaced in the Stull Lake belt, to crustally contaminated sources for continental alkaline volcanism at less than 2.71 Ga, when the Oxford Lake sedimentary subgroup was deposited, is interpreted to record the thrusting of the North Caribou craton beneath the Stull Lake–Edmund Lake belt. The driving force for this deformation was the oblique collision and regional metamorphism of the Northern Superior superterrane at ca. 2.72 to 2.71 Ga that is recorded in D_{2-4} structures. Erosion of both uplifted Northern Superior superterrane crust and the telescoped northern North Caribou margin provided sedimentary detritus for possible transtensional molasse basins along dextral strike-slip faults such as the WSWSZ, where mildly alkaline volcanic rocks erupted.

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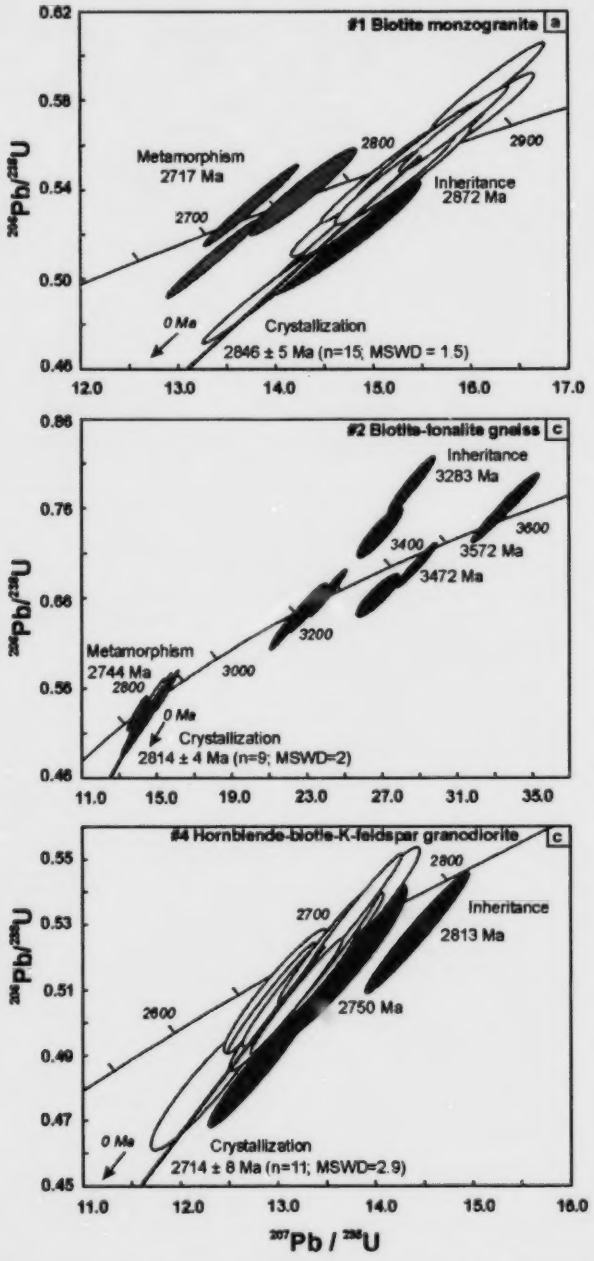

Figure GS-21-8: U-Pb concordia diagram of SHRIMP U-Pb zircon data from plutonic rocks in the Northern Superior superterrane. Sample locations are shown on Figure GS-21-1 (UTM coordinates in text). In all three diagrams, the crystallization age is given for a regression forced through the origin, and the number of analyses (open ellipse) and MSWD are given in parentheses. Filled (black) ellipses in all three diagrams are inherited zircon cores with selected $^{207}Pb/^{206}Pb$ ages. Filled (grey) ellipses in (a) and (b) are $^{207}Pb/^{206}Pb$ ages of metamorphic zircon overgrowths.

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GS-22 GEOLOGICAL INVESTIGATIONS IN THE KNEE LAKE BELT (PARTS OF NTS 53L)

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Corkery, M.T., Cameron, H.D.M., Lin, S., Skulski, T., Whalen, J.B. and Stern, R.A. 2000: Geological investigations in the Knee Lake belt (parts of NTS 53L); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 129-136.

SUMMARY

Mapping of the Knee Lake belt focused on the northeast in the Gods Lake area in 2000. Results include identification of two distinct, opposing, facing sequences within the Hayes River Group metavolcanic rocks, separated by a northeast-trending dextral fault with north side up, exposing amphibolite-grade rocks on the northwest and greenschist-grade rocks to the southeast. Geochronological studies yielded several important ages: 1) the Oxford Lake Group volcanic subgroup has a 2720 Ma depositional age; and 2) the Oxford Lake Group sedimentary subgroup has a broad range of detrital zircon ages ranging from 2711 to 3647 Ma. Possible sources of ancient ca. 3.6 Ga zircons occur in fault-bounded blocks along the northern margin of the northern Superior Province. Geochronological data from the Knee Lake segment date Hayes River volcanism at 2830 Ma, and sensitive high-resolution ion microprobe (SHRIMP) data on detrital zircons indicate that turbiditic sandstone is synvolcanic and volcanic-derived. Folding of the Hayes River Group predates deposition of coarsening-upward clastic sedimentary rocks and iron-formation, the former containing 2937 to 2822 Ma detrital zircons. The Oxford Lake volcanic subgroup was extruded at 2722 Ma in southern Knee Lake and was followed by deposition of fluvial-clastic, sedimentary subgroup sandstone with detrital zircons ranging in age from 2798 to 2707 Ma.

INTRODUCTION

The bedrock mapping projects of Manitoba's Northern Superior program are primarily designed to better understand the volcanic, structural and tectonic evolution of greenstone belts in the Manitoba portion of the Sachigo Subprovince–Gods Lake Domain. A secondary goal is to place the supracrustal belts in a regional context with the internal and surrounding plutonic and gneiss complexes. Central to this objective is an understanding of the largest contiguous greenstone belt in the Sachigo, the Knee Lake belt (Fig. GS-22-1). This report summarizes new mapping and laboratory results stemming from stratigraphic, geochemical, structural and geochronological studies of this belt (Syme et al., 1997, 1998; Corkery et al., 1999) being conducted as part of the Western Superior NATMAP Project. Preliminary geological maps for Gods Lake (Corkery and Cameron, 2000) cover the areas mapped in the Knee Lake belt in 1999 and 2000.

Mapping at a scale of 1:50 000 in the Elk Island area in 2000 extends the study of the supracrustal rocks in the Gods Lake area to the northeast of Gods Lake Narrows (Fig. GS-22-2). Existing 1:50 000 maps (Barry, 1961; Gilbert, 1985) formed an excellent base within which to integrate new observations.

A unique feature of Gods Lake is that it extends across two greenstone belts (Fig. GS-22-1), thereby providing a continuous section through plutonic and gneiss domains, both within and between the Knee Lake and Munroe Lake belts. A co-operative Manitoba Geological Survey (MGS)–Geological Survey of Canada (GSC) mapping and sampling program was conducted in all major plutonic and gneiss complexes on Gods Lake. This program is expected to provide important data for the Knee Lake belt project and also for a concurrent regional geochemical, isotopic and geochronological transect. The regional transect will focus on development and chronology of plutonic and gneissic terranes in the northwestern Superior Province.

The results of age determinations from samples collected during the past four years of mapping in the Knee lake belt are summarized. In past mapping programs, definition of the major assemblages was based primarily on field relationships. These were not always enough. The use of precision geochronology and geochemistry has provided a new discrimination tool.

Tectonostratigraphy

Supracrustal rocks in the Knee Lake belt have previously been assigned (Gilbert, 1985 and references therein) to two principal stratigraphic entities, the Hayes River Group (HRG) and the Oxford Lake Group (OLG). These groups are the principal focus of our current work, in that they have more potential to host gold and base-metal deposits than the voluminous granitoid terranes in the Superior Province.

The HRG (ca. 2830 Ma at Knee Lake; D. Davis, pers. comm., 1986) is a mainly volcanic sequence dominated by pillowed basalt and related gabbro, minor intermediate to felsic volcanic rocks, and minor volcanogenic sedimentary rocks. In the southeastern Gods Lake area, both the base and the top of the HRG have been intruded by tonalitic to granitic plutons and related gneissic units of the Bayly Lake complex (2783–2730 Ma; D. Davis, pers. comm., 1986).

The OLG (ca. 2706 Ma at Oxford Lake; D. Davis, pers. comm., 1986) is a largely sedimentary succession, up to 12 km thick, which lies unconformably on HRG volcanic rocks at Gods Lake (Gilbert, 1985). It consists of a lower, dominantly 'volcanic' subgroup of limited extent and an overlying, more extensive sedimentary subgroup. The OLG extends along the southern margin of the belt from Oxford Lake to Magill Lake, a distance of 40 km (Gilbert, 1985; Manitoba Energy and Mines, 1987), and discontinuously to the Gods Lake Narrows area. Volcanic rocks in the lower subgroup are shoshonitic to calc-alkalic in character (Hubregtse, 1976; Brooks et al., 1982; Gilbert, 1985). They include fragmental and flow rocks but are dominated by conglomeratic epilastic rocks (Syme et al., 1997). The overlying sedimentary subgroup is dominated by units deposited in a subaerial to locally shallow-marine environment.

ELK ISLAND AND CENTRAL GODS LAKE

Previous Work

Work in the Gods Lake area prior to 1970 was conducted by Wright (1931), Dix (1947) and Barry (1961). The Gods Lake Narrows map sheet by Barry (1961) was a particularly useful geological base for our investigations in the central part of Gods Lake.

The Gods Lake area was remapped in the early 1970s as part of the 'Greenstone Project' (bibliography in Gilbert, 1985). The Gods Lake Narrows area is covered by two three-colour 1:50 000 scale maps (GR83-1-12, -13) and one black and white preliminary map (1973H-12). A review of the exploration history of the area up to 1977 is contained in Southard (1977).

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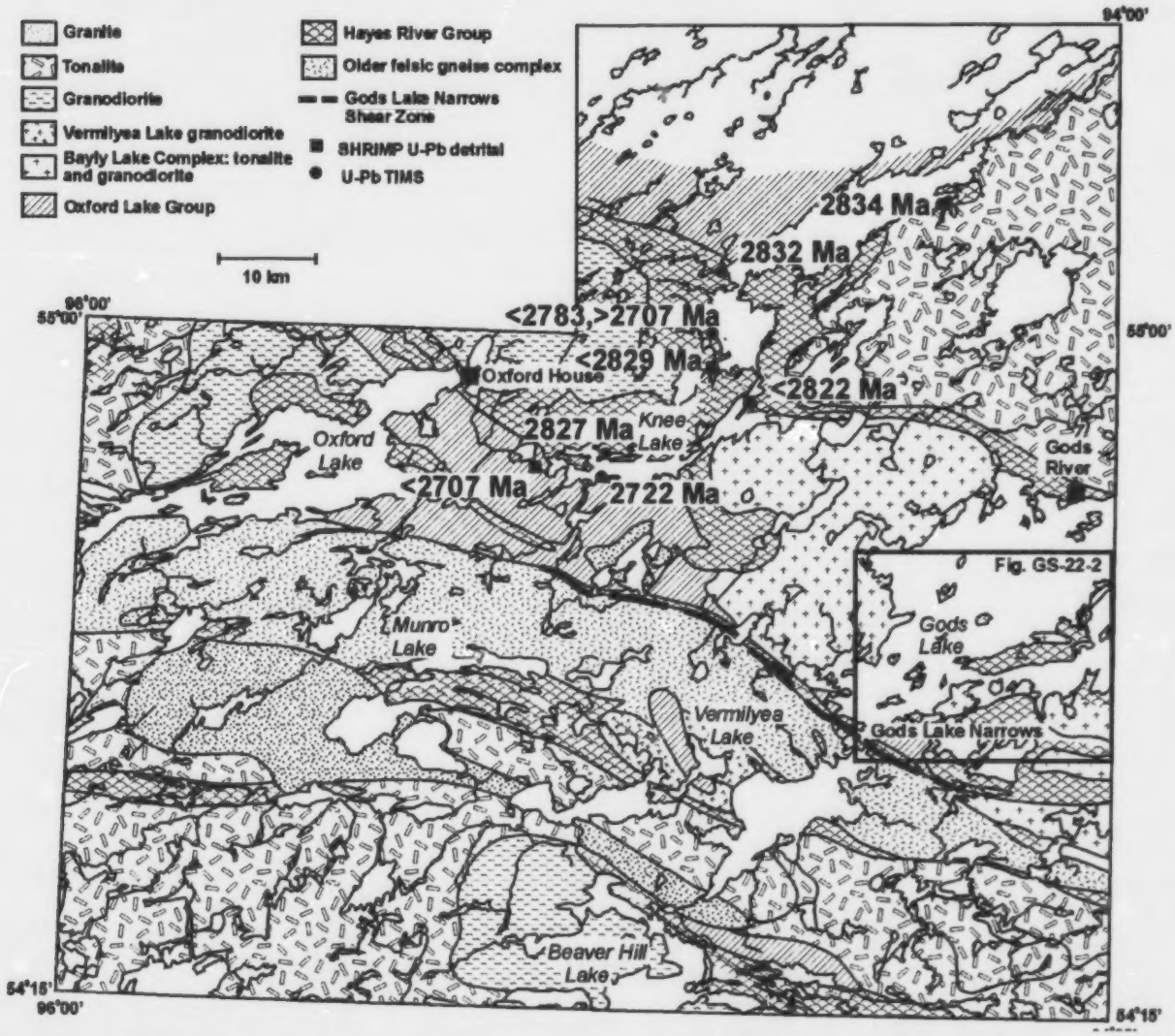

Figure GS-22-1: Regional geological setting and major subdivisions in the Gods Lake region. U-Pb zircon geochronology from this study.

Regional Setting

Northeast of Gods Lake Narrows, HRG mafic volcanic rocks form a homoclinal, northwest-facing, 2 to 3 km thick sequence that extends the full length of Elk Island (Fig. GS-22-2). The sequence is truncated to the south at an angular unconformity by east-striking and south-facing conglomerate rocks of the Oxford Lake sedimentary subgroup. To the southeast, HRG rocks are intruded by granite of the Bayly Lake complex and, on the northwest, the HRG is truncated by a northeast-trending fault zone. Northwest of the fault, a southeast-facing heterolithic sequence of HRG rocks consists of a lower, mainly sedimentary part and a predominantly basaltic upper part.

The OLG, south of the unconformity, is contained in a southeast-trending, southwest-facing belt up to 5 km wide (Fig. GS-18-2). The sedimentary subgroup directly overlying the HRG forms a 1 km thick, largely subaerial succession, which is in fault contact to the south with a second south-facing sequence dominated by a submarine succession of fine-grained sedimentary rocks of Oxford Lake volcanic subgroup affinity.

The southern boundary of the Knee Lake belt is marked by a major, 1 to 1.5 km wide mylonite that extends through Gods Lake Narrows. South of this feature, a 10 km thick, older felsic gneiss complex (U-Pb age of 2883 Ma; Davis pers. comm., 1986) separates the Knee Lake belt from several thin slivers of supracrustal rocks that form the eastern extension of the Munroe Lake belt.

Hayes River Group

Two distinct sequences previously reported as HRG (Gilbert, 1985) occur in the Elk Island area:

1) The Elk Island sequence is a homoclinal, northwest-facing sequence of HRG mafic volcanic rocks that represents a continuation of the basaltic section in the Gods Lake Narrows area (Corkery et al., 1999).
2) The Billy Hole Island sequence is a southeast-facing sequence of coarsening-upward metasedimentary and metavolcanic rocks that occurs directly northwest of the basaltic rocks (Fig. GS-22-2).

Elk Island sequence

In the northwest-facing basalt sequence at Elk Island, primary structures and textures are remarkably well preserved at greenschist-facies metamorphic grade throughout most of the basalt-dominated section (Fig. GS-22-3). There is little or no penetrative fabric in most outcrops. Near fault zones, a narrow strain gradient is exhibited as a relatively abrupt increase in a penetrative fracture cleavage and shear

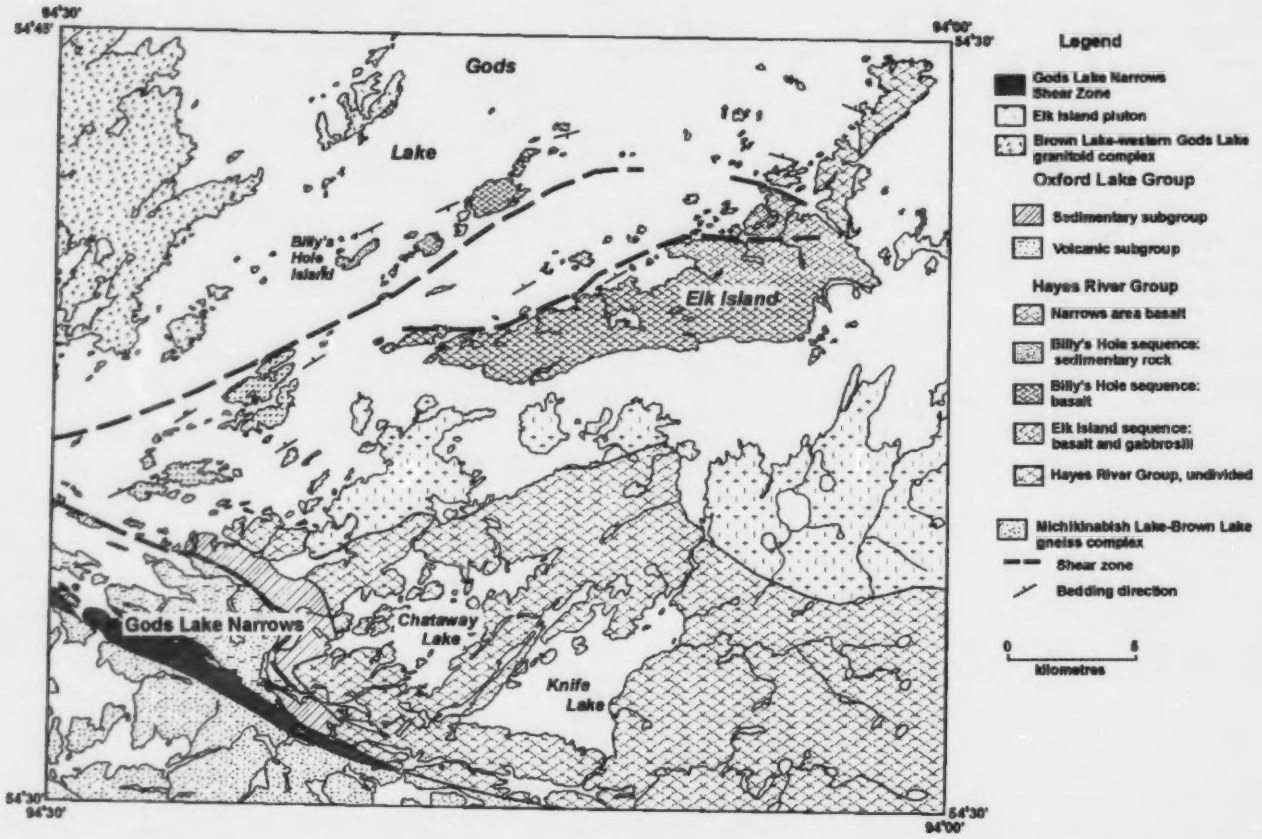

*Figure GS-22-2: General geology of south and central Gods Lake (*modified from *Manitoba Energy and Mines, 1987).*

fabric grading into mylonite. The two major shear zones within the section generally strike northeasterly, with several offshoots that are conformable with the east-striking stratigraphy (Fig. GS-22-2).

The Elk Island sequence is dominated by massive and pillowed, aphyric to sparsely plagioclase-phyric flows and subordinate amoeboid pillow and flow-top breccia. These flows have characteristic beige to medium green weathering colours and are pale to medium green on fresh surfaces. Pillowed flows display thin selvages and contain thick interpillow hyaloclastite or interpillow chert. Pillows typically have vesicular margins and commonly contain 1 mm by 2 to 5 mm radial pipe-vesicles. Numerous synvolcanic gabbro sills and dykes intrude the sequence; these range from minor veins to sills over 100 m thick. Some gabbroic units are associated with thick amoeboid pillow and flow-top breccia units (Fig. GS-22-4) and may represent thick, ponded basalt flows. One such unit extends along the north shore of Elk Island for approximately 4 km in the area southwest of the old Gods Lake gold mine headframe (Fig. GS-22-2).

Billy's Hole Island sequence

A southeast-facing sedimentary and volcanic section, previously mapped as Hayes River Group, occurs northeast of the Elk Island

Figure GS-22-3: Well preserved pillowed basalt flow, Elk Island area.

Figure GS-22-4: Thick basalt flow-top breccia with amoeboid pillows, Elk Island area.

basaltic sequence in the vicinity of Billy's Hole Island (Fig. GS-22-2). The volcano-sedimentary section is intruded on the north and west by granite of the Bayly Lake complex (2730 Ma; D. Davis, pers. comm., 1986); the section is bounded on the south by a northeast-trending fault (Fig. GS-22-2). Primary structures and textures are variably preserved at amphibolite facies throughout.

The volcano-sedimentary section is composed of a number of distinct stratigraphic units. These can be broadly grouped into a lower sediment-dominated sequence that is gradational with a mixed epiclastic and volcaniclastic member and an uppermost pillowed mafic volcanic member. The lowermost part of the section is dominated by fine-grained sedimentary rocks: feldspathic greywacke sandstone, siltstone and argillite. These grade upward into medium- to coarse-grained sandstone with interbeds of polymictic pebble-cobble, matrix-supported conglomerate (Fig. GS-22-5) with a significant felsic and mafic volcanic detrital component. Rhyolite breccia, massive rhyolite, and rare massive and pillowed basalt flows are interlayered with the sedimentary rocks in the central portion of the section. These are, in turn, overlain by a 1 to 2 km thick sequence of aphyric and plagioclase-phyric pillowed basalt that constitutes the upper part of the section.

Oxford Lake Group

Both the volcanic and sedimentary subgroups of the OLG are exposed at Gods Lake Narrows. The volcanic subgroup forms a 3 to 4 km thick, fault-bounded panel (Fig. GS-22-2). Although most rocks in this panel are highly strained, primary structures and textures (including graded bedding) in the sedimentary rocks indicate this to be a southward-younging sequence.

In Gods Lake Narrows, the OLG sedimentary subgroup is a simple, 1 to 1.5 km thick, continuous, south-facing sequence of well preserved polymictic conglomerate. A transect through the eastern end of the unit revealed no indication of either fining or coarsening upward. The OLG is not well represented in the Elk Island area; however, several spectacular exposures of the OLG sedimentary subgroup, with an unconformable relationship to the Billy's Hole Island sequence, are exposed (Fig. GS-22-2). Polymictic framework boulder conglomerate, of apparent greenschist-facies metamorphic grade, rests unconformably on an irregular depositional surface (Fig. GS-22-6) of amphibolite-facies grade basalt.

Granitoid Rocks

Most of the granitoid rocks in the Knee Lake–Gods Lake area were interpreted as pre-Oxford Lake Group and assigned to the 'Bayly Lake complex' (Gilbert, 1985). Lack of high-resolution geochemical and geochronological discrimination tools at the time prevented a more sophisticated investigation. In the present study, all major units previously identified have been examined and sampled.

Figure GS-22-5: Polymictic matrix-supported conglomerate, central portion of Billy's Hole Island sequence.

Figure GS-22-6: Polymictic conglomerate of the OLG on an unconformity at the top of the Billy's Hole Island sequence.

Brown Lake–western Gods Lake granitoid complex and Bayly Bay pluton

The Brown Lake–western Gods Lake granitoid complex that dominates northwestern Gods Lake is intruded by the Bayly Bay pluton (Fig. GS-22-2; Gilbert, 1985). However, the granitoid complex and the Bayly Bay pluton both contain xenoliths of an older group of well to strongly foliated to gneissic, biotite-hornblende tonalite to granodiorite gneiss. These rocks are intruded by two younger phases of more homogeneous tonalite to granodiorite (Gilbert, 1985) that were broadly subdivided into 1) fine- to medium-grained biotite±hornblende tonalite; and 2) younger, medium- to coarse-grained tonalite to granodiorite. Both tectonic and intrusive contact relationships have been observed between the Brown Lake–western Gods Lake granitoid complex rocks and the HRG. Geochemical, Nd isotopic and U-Pb geochronological work in progress will determine their tectonomagmatic affinity and should constrain whether this suite is composed of cogenetic granitoid rocks.

Elk Island pluton

Younger, massive to slightly foliated, biotite monzogranite to granodiorite forms an intrusion south and west of Elk Island (Fig. GS-22-2). The intrusion appears to be late, high level and possibly postorogenic or anorogenic. A shallow level of emplacement and erosion is indicated by the presence of spectacular tuffisite breccia in granite shoreline exposures southeast of Elk Island (Fig. GS-22-7). Tuffisite represents a degassing breccia formed during the escape of magma and magmatic gas through congealed upper parts of shallow (approx. 2 km deep) intrusions, at first passively but later with increasingly explosive effects (Hughes, 1971). The Elk Island pluton is probably comagmatic with earlier portions of the OLG, based on 1) the geochronology summarized in the text for the Oxford Lake Group in Manitoba (2720–2729 Ma), plus the age range reported for the Bayley Lake complex (2783–2730 Ma); 2) the 'intracontinental' character of the OLG, as evidenced by calc-alkalic to shoshonitic volcanic rocks and dominantly subaerial sedimentary rocks; and 3) the evolved high-level features of the granitoid intrusions. Granitoid geochemical work in progress may help in understanding the tectonic regime in which the OLG was deposited. These intrusions have a bearing on both the pre- and postemplacement uplift history of the Knee Lake belt.

Michikinabish Lake–Brown Lake gneiss complex

South of the Gods Lake Narrows Shear Zone (GLNSZ), there is a 10 km wide belt of biotite-hornblende tonalitic to quartz dioritic gneiss, a sample of which yielded a U-Pb age of ca. 2.9 Ga (D. Davis, pers. comm., 1986). These rocks display lithological features typical of Archean trondhjemite-tonalite-granodiorite (TTG) suites. The Nd isotopic work in progress on this suite should help constrain whether the GLNSZ juxtaposes different crustal blocks or different crustal levels of the same block.

Figure GS-22-7: Tuffisite in the Elk Island granodiorite.

GEOCHRONOLOGICAL RESULTS FROM THE KNEE LAKE BELT

Eight new U-Pb zircon ages were obtained from the Knee Lake area (northwest of Gods Lake) in order to unravel the timing of volcanism and sedimentation in well studied sections of the Hayes River and Oxford Lake groups (cf. Syme et al., 1997, 1998). Initial results of this geochronological investigation are reported here to provide a regional context within which to interpret both the tectonostratigraphy of the Gods Lake area and the two new U-Pb zircon ages from the Gods Lake Narrows segment of the Knee Lake greenstone belt (D. Davis, pers. comm., 2000). Extraction, dissolution and U-Pb analysis of zircon and titanite by thermal-ionization mass spectrometry (TIMS) was carried out on volcanic and plutonic rocks at the Geological Survey of Canada (cf. Percival and Skulski, 2000, and references therein). Detrital zircons from sedimentary rocks were analyzed with the GSC's SHRIMP II facility using extraction and analytical methods described by Stern (1997).

Three samples of HRG felsic volcanic rocks were dated in the Knee Lake area (Fig. GS-22-8). A sample of quartz-phyric, rhyolite lapilli tuff, locally overlain by resedimented volcaniclastic rocks, was collected from the south shore of northern Knee Lake (lat. 55°07'00"N, long. 94°31'37.21"W; Fig. GS-22-1). Four single-zircon fractions were analyzed and three of these yielded a U-Pb age of 2834 +3/-2 Ma (Fig. GS-22-8a). A sample of resedimented felsic lapilli tuff that contained clasts of siltstone, abundant quartz fragments and lithic fragments (including quartz-phyric rhyolite), was collected from the northwestern shore of northern Knee Lake (lat. 55°04'2.70"N, long. 94°45'13.95"W; Fig. GS-22-1). Three single-grain and one multigrain (n=4) zircon fractions were analyzed, and two single-zircon fractions yielded a U-Pb age of 2832 +2/-1.5 Ma (Fig. GS-22-8b). A third sample of flow-banded rhyolite from the HRG was collected from the north shore of Pain Killer Bay, in southern Knee Lake (lat. 54°51'00"N, long. 95°9'00"W; Fig. GS-22-1). This rhyolite occupies the middle formation of the Pain Killer Bay complex (Syme et al., 1997), which consists of a high-level porphyritic felsic intrusion overlain successively by a lower formation of siltstone and feldspathic sandstone, a middle formation of massive and fragmental rhyolite flows, and an upper formation of felsic volcaniclastic-sedimentary and pyroclastic rocks. Five multigrain zircon fractions yielded a U-Pb age of 2827 +5/-4 Ma (Fig. GS-22-8c). These data demonstrate a relatively consistent pattern (within error) for HRG calc-alkaline felsic volcanism at ca. 2830 Ma.

Detrital zircon ages from a synvolcanic greywacke in the HRG provide further constraints on the HRG sources, and the age and duration of Hayes River volcanism. A sample of feldspathic wacke from a turbiditic succession of quartz-poor sandstone and siltstone that overlies the northwestern flank of the Pain Killer Bay rhyolite complex (2827 +5/-4 Ma, above) was collected from the western shore of central Knee Lake (lat. 54°47'48"N, long. 94°47'48"W; Fig. GS-22-1). Twenty detrital zircons from this sample were analyzed with the SHRIMP and yielded a narrow range of ages, from 2853 ± 10 Ma to 2829 ± 9 Ma; the latter represents the average of four analyses of a single grain and is the maximum age of sedimentation (Fig. GS-22-9a). This maximum age constraint is within the margin of error for the age of the Pain Killer Bay rhyolite, and this result is consistent with the notion that sedimentation was approximately contemporaneous with felsic volcanism (Syme et al., 1997, and references therein).

In central Knee Lake, the Opischikona Narrows sedimentary rocks are separated by an angular unconformity from underlying, previously folded volcanic rocks of the HRG (Syme et al., 1997, 1998). This coarsening-upward sedimentary sequence includes a lower unit of chert-magnetite iron-formation and argillite overlain by greywacke, pebbly sandstone and conglomerate. A sample of greywacke from a small island in Opischikona Narrows, central Knee Lake (lat. 55°01'32.43"N, long. 94°47'04.65"W; Fig. GS-22-1) was collected for detrital zircon dating with the SHRIMP. Seventeen detrital zircon grains were analyzed and found to encompass an age range from 2937 ± 12 Ma to 2822 +18/-15

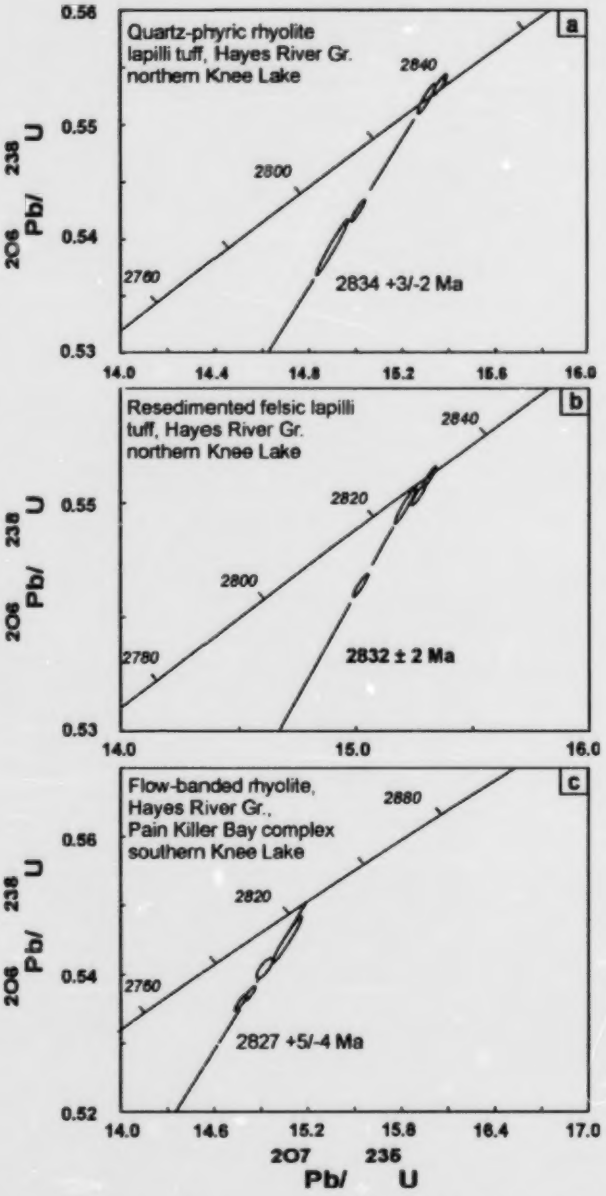

Figure GS-22-8: U-Pb zircon geochronology of the Hayes River Group, Knee Lake segment of the Knee Lake greenstone belt: a) quartz-phyric rhyolite lapilli tuff from northern Knee Lake; b) resedimented felsic lapilli tuff from northern Knee Lake; and c) flow-banded rhyolite from Painkiller Bay.

Ma, the latter being an average of four points on one grain and representing the maximum age of greywacke deposition (Fig. GS-22-9b). The detrital zircon ages reflect sediment sources not only in the immediately underlying Hayes River Group, but from older Mesoarchean rocks as well. Some of the older ca. 2.93 Ga detrital zircons are large (200 µm in length), slightly rounded grains typical of plutonic rocks and may reflect uplift of crystalline basement. In order to constrain a minimum age of sedimentation, a sample of a feldspar porphyry sill that cuts the Opischikona Narrows sedimentary rocks was sampled from central Knee Lake (lat. 55°01'32.43"N, long. 94°47'04.65"W; Fig. GS-22-1). Two single-grain zircon fractions and one multigrain fraction (n=2) were analyzed by TIMS and yielded $^{207}Pb/^{206}Pb$ ages ranging from 2888 Ma

to 2783 Ma, the latter providing a maximum age for the sill (Fig. GS-22-9c). Two multigrain titanite fractions were analyzed to obtain a minimum age for the sill; these yielded ages of 2709 ± 12 Ma and 2707 ± 17 Ma, the latter being less discordant and providing a preliminary minimum age for the sill. Opischikona Narrows sedimentary rocks were therefore deposited after 2822 Ma (youngest detrital zircon) and before 2707 Ma (titanite cooling or metamorphic age of crosscutting sill). These data allow for the interpretation that the Opischikona sedimentary rocks represent an outlier of Oxford Lake Group. Further work on the feldspar porphyry sill will refine testing of this hypothesis.

A sample of dacitic lapilli tuff from the volcanic subgroup of the OLG, exposed on an island in southern Knee Lake (lat. 54°50'36"N, long. 94°57'48"W; Fig. GS-22-1), was dated to constrain the age of volcanism. This volcanic rock is part of a sequence of intermediate shoshonitic to calc-alkaline volcaniclastic rocks and mafic lavas on southern Knee Lake. Two multigrain fractions and one single-grain zircon fraction yielded a U-Pb age of 2722 ± 3 Ma (Fig. GS-22-10a). A sample of dacite lapilli tuff located between GLNSZ and conglomerate of the Oxford Lake sedimentary subgroup (lat. 54°34'58"N, long. 94°30'00"W; Fig. GS-22-1), yielded a U-Pb zircon age of 2719.9 ± 1.4 Ma (D. Davis, pers. comm., 2000). Elsewhere in northern Manitoba and Ontario, the Oxford Lake volcanic subgroup has yielded U-Pb zircon ages of 2726 ± 2 Ma on Little Stull Lake (Corkery and Skulski, 1998), greater than 2728 ± 2 Ma on Edmund Lake in the east (Corkery and Heaman, 1998), and 2729 ± 3 Ma for the Gunpoint Group on Cross Lake in the west (Corkery et al., 1992).

A sample of fluvial, crossbedded feldspathic wacke with distinctive heavy-mineral trails was collected from the sedimentary subgroup of the Oxford Lake Group at Trout Falls in the southern Knee Lake area (lat. 54°51'21"N, long. 95°7'30"W; Fig. GS-22-1) for dating of detrital zircons. Twenty-five detrital zircon grains were analyzed with the SHRIMP and yielded ages ranging from 2798 ± 12 Ma to 2707 +9/-8 Ma, the latter representing a maximum age of sedimentation derived by averaging five point analyses on a single grain. Detrital zircons in this sample have ages 1) slightly younger than Hayes River volcanism (ca. 2830 Ma, see above), 2) typical of the volcanic subgroup of the Oxford Lake group (ca. 2730 Ma), and 3) typical of felsic volcanism recorded in the OLG on Oxford Lake (2706 +4/-2 Ma; Syme et al. 1993). A sample of crossbedded sandstone from near the base of the polymictic conglomerate-dominated sedimentary subgroup of the OLG south of the GLNSZ was collected for U-Pb dating of detrital zircons by TIMS (D. Davis, pers. comm., 2000). Five detrital grains yielded $^{207}Pb/^{206}Pb$ ages

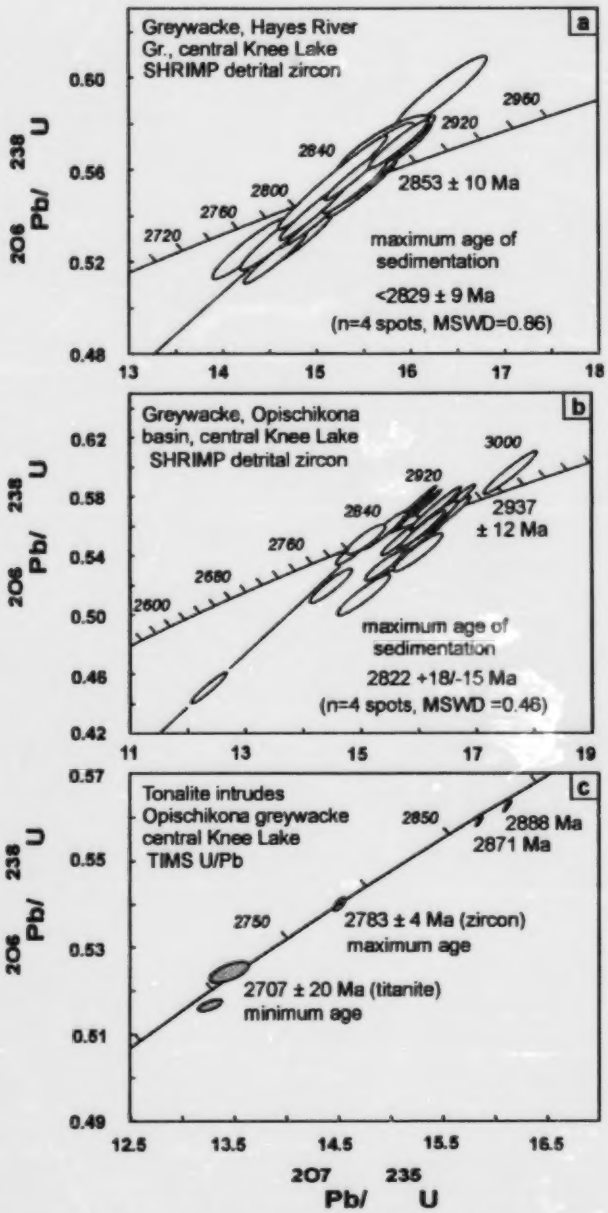
Greywacke, Hayes River Gr., central Knee Lake SHRIMP detrital zircon
a
2853 ± 10 Ma
maximum age of sedimentation
<2829 ± 9 Ma
(n=4 spots, MSWD=0.86)
Greywacke, Opischikona basin, central Knee Lake SHRIMP detrital zircon
b
2937 ± 12 Ma
maximum age of sedimentation
2822 +18/-15 Ma
(n=4 spots, MSWD =0.46)
Tonalite intrudes Opischikona greywacke central Knee Lake TIMS U/Pb
c
2888 Ma
2871 Ma
2783 ± 4 Ma (zircon)
maximum age
2707 ± 20 Ma (titanite)
minimum age
206 Pb/ 238 U
207 Pb/ 235 U

Figure GS-22-9: U-Pb geochronology of a) Hayes River Group greywacke (SHRIMP detrital zircon); b) Opischikona basin greywacke (SHRIMP detrital zircon); and c) tonalite that cuts the Opischikona basin (U-Pb TIMS zircon and titanite).

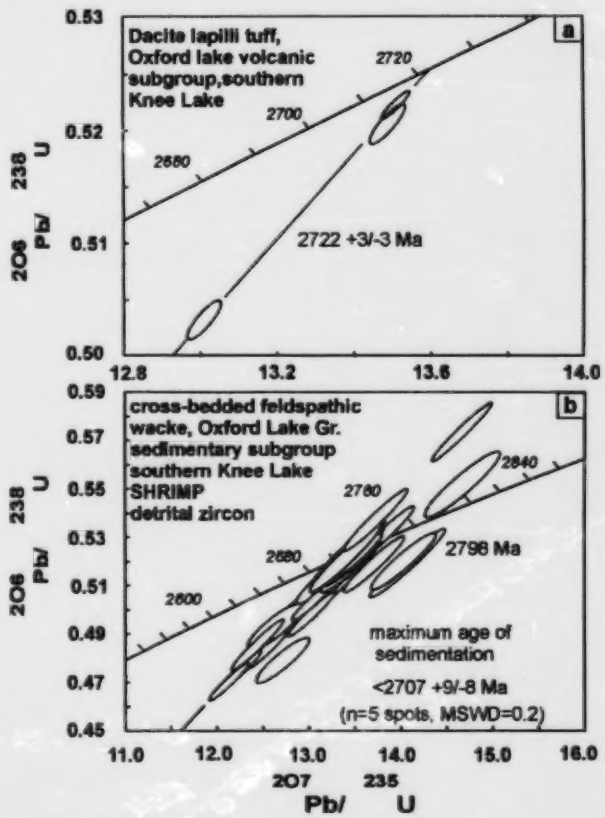
Dacite lapilli tuff, Oxford lake volcanic subgroup,southern Knee Lake
a
2722 +3/-3 Ma
cross-bedded feldspathic wacke, Oxford Lake Gr. sedimentary subgroup southern Knee Lake SHRIMP detrital zircon
b
2798 Ma
maximum age of sedimentation
<2707 +9/-8 Ma
(n=5 spots, MSWD=0.2)
206 Pb/ 238 U
207 Pb/ 235 U

Figure GS-22-10: U-Pb zircon geochronology of the Oxford Lake Group a) dacite lapilli tuff volcanic subgroup (U-Pb TIMS zircon); and b) crossbedded feldspathic wacke, sedimentary subgroup (SHRIMP detrital zircon).

of 3647 ± 2 Ma, 2846 ± 1 Ma, 2827 ± 1 Ma, 2721 ± 2 Ma and 2711 ± 2 Ma, the latter representing the maximum age of sedimentation. Possible sources of ancient ca. 3.6 Ga zircons occur in fault-bounded blocks along the northern margin of the northern Superior Province. At Assean Lake in northwestern Manitoba, a 3.54 Ga (U-Pb zircon age) tonalite gneiss occurs within a widespread zone of granitoid and gneissic rocks with Nd model ages greater than 3.5 Ga (Böhm et al., 2000). In northern Ontario, north of the North Kenyon Fault, granitoid rocks have greater than 3.0 to 3.5 Ga Nd model ages, and 3.57 Ga (SHRIMP) inherited zircon cores occur in 2.81 Ga tonalite gneiss (Skulski et al., GS-21, this volume). Crust of this age is not known south of these two areas, and the first evidence of recycling of greater than 3.5 Ga zircons includes detrital zircon in less than 2.71 Ga Oxford Lake sedimentary subgroup rocks at Gods Lake Narrows (this study), and 3.5 Ga detrital zircon in the correlative (<2704 Ma) Cross Lake Group to the west. These young sedimentary sequences may provide a minimum age for late docking of this older crust with the Superior Province (Skulski et al., GS-21, this volume).

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SOUTHEAST MAX LAKE AREA (PARTS OF NTS 53L/5N, /6NW, /12SW)

by H.P. Gilbert

Gilbert, H.P. 2000: Southeast Max Lake area (parts of NTS 53L/5N, /6NW, /12SW); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 137-149.

SUMMARY

- The Ralph Anderson Lake greenstone belt is a branch of the Archean 'Max Lake belt', and extends from Logan Lake in the west to Aswapiswanan Lake in the east, a distance of approximately 50 km. The greenstone belt is essentially monoclinal, and consists mainly of arc-type mafic volcanic rocks, with subordinate felsic volcanic and sedimentary units. An early synclinal fold is located close to the north margin of the west part of the belt.
- The lower (southern) part of the Ralph Anderson Lake belt is intruded by an approximately 1.2 km thick gabbro sill. A series of skialithic enclaves in the gabbro, which constitute the lower part of the supracrustal sequence, consist of basalt, heterolithic debris flows and turbidite, and rhyolitic rocks. The overlying main basaltic sequence that constitutes the major part of the belt is divided into geochemically distinct, southern and northern volcanic suites by a conspicuous gabbro (Lavigne Lake gabbro), which extends for more than 24 km through the central part of the belt. A felsic volcanic-sedimentary formation (together with an associated subvolcanic porphyry sill) occurs close to the north (upper) margin of the belt.
- The mafic volcanic rocks, consisting mainly of pillowed aphyric basalt, are characterized by widespread sea-floor–type hydrothermal alteration. A prominent, garnetiferous, silicic alteration zone in the northwest part of the map area is coincident with an oxide-facies iron-formation, and is interpreted as a site of early hydrothermal alteration. Alteration associated with later deformation is also widespread in the map area, and is especially conspicuous at the south margin of the main basaltic sequence.
- Pyritic mineralization occurs in various stratigraphic and structural settings. The best prospects for base-metal exploration appear to be the garnetiferous alteration zone in the northwest part of the map area, and stratabound mineralization associated with chert-siltstone units at various stratigraphic levels within both the southern and northern volcanic suites. Sulphide mineralization is also common in magnetiferous iron-formation, which offers potential base-metal and precious-metal exploration targets. The contact between the volcanic terrane and gabbro at the south margin of the greenstone belt contains sporadic pyritic zones that may represent prospects for economic mineralization. Lastly, a gabbro-pyroxenite intrusion that extends through the central part of the Ralph Anderson Lake belt is currently being tested for possible platinum-group element (PGE) mineralization.

INTRODUCTION

The southeast Max Lake area is located 120 km east-northeast of Norway House, at the junction of the Archean Gods Lake and Molson Lake domains (Manitoba Energy and Mines, 1987). The map area is centred on a series of small lakes immediately southeast of Max Lake, and encompasses part of an approximately 3 km wide supracrustal belt (Ralph Anderson Lake belt) that extends from Logan Lake in the west to Aswapiswanan Lake in the east, a strike length of approximately 50 km (Fig. GS-23-1). This belt merges with the northeast-trending Max Lake belt in the west (Hubregtse, 1974); to the east, the belt is on strike (across a drift-covered area) with supracrustal rocks in the Goose Lake–Beaver Hill Lake area (Elbers and Marten, 1973), which are inferred to be laterally continuous with the Ralph Anderson Lake belt.

A two-year mapping program was initiated in 1999 to upgrade

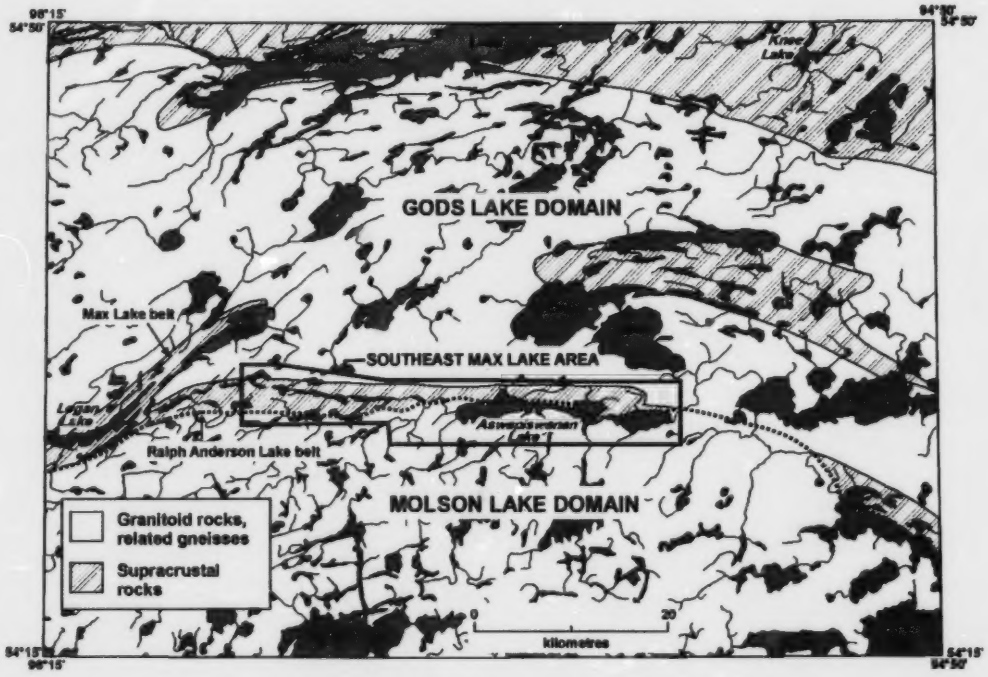

Figure GS-23-1: Regional geological map showing the location of the project area at the junction of the Gods Lake and Molson Lake domains.

previous 1:250 000 scale maps (Currie, 1961; Manitoba Energy and Mines, 1987) and to provide a detailed map as a base for possible further geological studies. Previous geological maps in the southeast Max Lake area show only rudimentary stratigraphic and structural information (Currie, 1961; Elbers, 1973a, b; Schledewitz, 1980). Forest fires in 1989 throughout most of the map area resulted in a significant increase in the amount and quality of bedrock exposure, and thus increased the scope for detailed mapping. In 1999, 1:20 000 scale mapping in the MacVicar Lake–Ralph Anderson Lake area focused on stratigraphy, structure and the geochemistry and alteration of volcanic rocks; a conspicuous 36 m wide, garnetiferous, silicified zone in the northwest part of the map area was the subject of particular attention (Gilbert, 1999a). Field mapping in the 2000 season extended the 1999 map area east to the Franklin Murray Lake–Aswapiswanan Lake area, and also farther west in the MacVicar Lake area.

The Ralph Anderson Lake belt consists mainly of basaltic pillow lavas with subordinate felsic volcanic rocks, heterolithic breccia and epiclastic deposits, flanked by granitoid rocks to the north and a 1.2 km wide gabbroic intrusion (McLeod Narrows gabbro) to the south (Fig. GS-23-2, -3; Table GS-23-1). The supracrustal rocks are provisionally equated with the volcanosedimentary 'Hayes River Group' in the Oxford Lake–Knee Lake–Gods Lake area northeast of Max Lake (Hubregtse, 1985a and references therein). The margins of the Ralph Anderson Lake belt are characterized by contact zones in which the supracrustal rocks are intruded by conformable units derived from the flanking plutonic terranes. The granitoid terrane north of the Ralph Anderson Lake belt consists of tonalitic to granodioritic intrusions and related gneissic units (Hubregtse, 1985b). The McLeod Narrows gabbro south of the belt contains volcanic and related sedimentary skialithic enclaves (up to 300 m thick) considered to be part of the supracrustal sequence, which is interpreted as essentially monoclinal. Abundant facing directions, indicated by pillow structure and sporadic top indicators in epiclastic rocks, are almost exclusively to the north. A synclinal fold near the north margin of the belt is indicated by a south-facing turbidite deposit in the west part of the project area. Whereas regional strike-parallel faults have not been identified within the greenstone belt, the effects of possible faulting (i.e. stratigraphic hiatus or repetition) cannot be discounted.

Fifty rock samples from volcanic and related intrusive map units were analyzed for major, trace and rare-earth elements. Mafic to felsic volcanic and associated intrusive rocks are geochemically akin to modern arc tholeiite. They display marked negative Nb anomalies and flat to negatively sloping rare-earth element (REE) profiles due to decoupling between large-ion lithophile elements (LILE) and high field-strength elements (HFSE); TiO_2 values are generally less than 1.0%. Interflow gabbroic sills are compositionally similar to mafic volcanic units. The mafic-ultramafic Lavigne Lake gabbro sill in the central part of the belt, which is relatively less fractionated than the synvolcanic gabbro bodies, may be penecontemporaneous with volcanism and represent a more primitive fraction of the source magma for the volcanic rocks.

This report provides a summary of the geology of the Franklin Murray Lake–Aswapiswanan Lake area mapped this year (Fig. GS-23-4), as well as new data in the vicinity of MacVicar Lake (west part of the project area; Fig. GS-23-5), where mapping was initiated in 1999. The reader should refer to Table GS-23-1 for abbreviated descriptions of all subunits discussed in the text. Stratigraphic details in the MacVicar Lake–Ralph Anderson Lake area were published in Gilbert (1999a). An open file report, scheduled for publication in late 2000 or early 2001, will contain a geological description of the entire project area, together with a synthesis of the geochemistry of the volcanic rocks, the tectonic and metamorphic history, and a discussion of the economic potential of the belt.

STRATIGRAPHY OF THE FRANKLIN MURRAY LAKE–ASWAPISWANAN LAKE AREA

The stratigraphic sequence defined in the west part of the Ralph Anderson Lake greenstone belt (Gilbert, 1999a) extends eastward to the vicinity of Franklin Murray Lake. Farther to the east, in the Aswapiswanan Lake area, stratigraphic details are incomplete due to more intense deformation and metamorphism, the occurrence of pervasive intrusive rocks and an extensive cover of glacial drift. All volcanic supracrustal rocks in the Aswapiswanan Lake area occur south of an east-trending gabbro/pyroxenite sill that is interpreted as the on-strike equivalent of the Lavigne Lake gabbro, and thus constitute the 'south part' of the volcanic section (Gilbert, 1999a). Details of the 'north part' of the sequence (Gilbert, 1999a) are not known due to lack of exposure; supracrustal rocks in that area have either been covered by glacial drift and swamp, or they have been removed due to assimilation by granitoid intrusions that are part of an extensive plutonic terrane to the north of the greenstone belt. The greenstone belt is increasingly more attenuated and disrupted by granitoid rocks toward the east, in the Franklin Murray Lake–Aswapiswanan Lake area. At the south margin of the belt, strongly deformed basalt and laminated amphibolite have been folded and subsequently incorporated within granitic to quartz dioritic rocks. In that locality, a distinctive, plagioclase-phyric tonalite is interlayered with basalt-derived amphibolite and granitoid rocks in a gneissic zone (0.5–1.0 km wide) between the greenstone belt to the north and granodioritic rocks to the south.

The following section describes the main stratigraphic components of the Franklin Murray Lake–Aswapiswanan Lake area; equivalent rock units in the MacVicar Lake–Ralph Anderson Lake area to the west, which are relatively less deformed and better preserved, are described in Gilbert (1999a).

Basalt, Related Fragmental and Intrusive Rocks; Derived Laminated Amphibolite, Schist and Gneiss (Unit 1)

Aphyric basalt (subunit 1a) is the principal rock type in the Ralph Anderson Lake greenstone belt at the west end of Aswapiswanan Lake, where mafic flows are intercalated with gabbro and laminated amphibolite in a section up to 1.4 km wide. North-facing pillows were observed at one locality within this section, but the mafic volcanic rocks are generally too deformed for tops to be determined; pillow elongations are typically 8:1 to 15:1. Amygdaloidal zones occur sporadically (plagioclase±quartz, or carbonate amygdales); epidote alteration is common, either in the cores or toward the margins of pillows; silicic alteration at pillow rims is subordinate. Basalt in the Franklin Murray Lake–Aswapiswanan Lake area is intimately intercalated with subordinate synvolcanic gabbro units up to 20 m thick (subunit 1c). The mafic sills are largely massive and equigranular, but a plagioclase-megaphyric phase (subunit 1d), with equant feldspars up to 2.5 cm across, occurs at the west extremity of Aswapiswanan Lake.

In strongly deformed zones, basalt is locally altered to gneiss (subunit 1e) with fine (1–15 mm) laminae that are variously hornblendic (±chlorite), epidotic or quartzofeldspathic. The tectonometamorphic lamination is due to the attenuation of previous epidotic or silicic alteration domains and chloritic selvages in the pillowed flows. Stringers and aggregates of epidote (±quartz±carbonate±garnet) postdate the lamination. Elsewhere, where metamorphism was not accompanied by strong deformation, mafic volcanic rocks locally contain diffuse, recrystallized domains of fine- to medium-grained plagioclase+hornblende.

Narrow (10–40 cm thick) alteration zones with surficial iron staining and silicification (±garnet porphyroblastesis) occur in several mafic flow units. Similar gossan zones occur sporadically within or at the margins of synvolcanic gabbro sills (subunit 1c). These zones contain disseminated pyrite and pyritic stringers and local trace to minor concentrations of Cu, Zn, Ni, Cr and, at one locality, Au (*see* 'Economic Geology' section).

In the central and east parts of Aswapiswanan Lake, bedrock exposure is generally poor, except for a basaltic section, more than 1.2 km wide, that extends north from the eastern lobe of the lake. Scattered outcrops along the south shore of the lake display sporadic blocks or

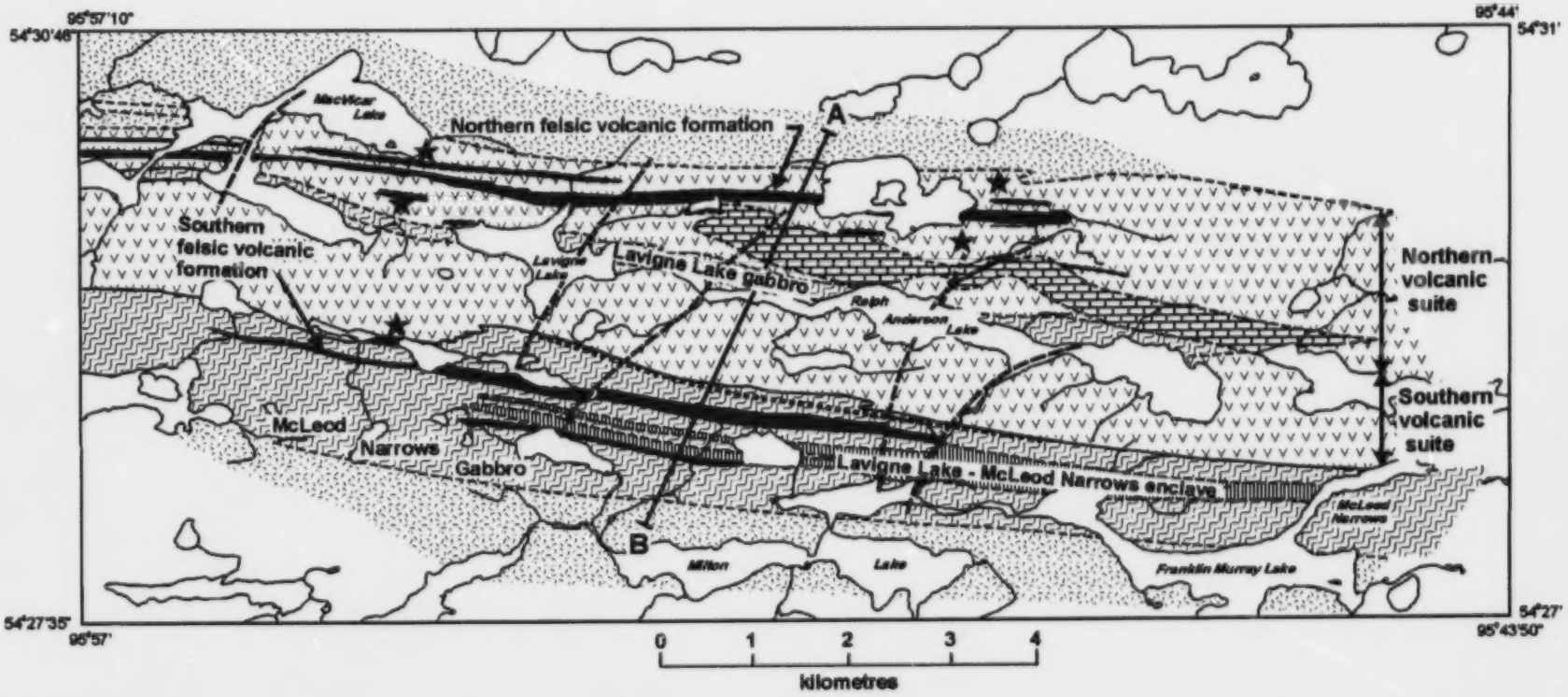

Figure GS-23-2: Simplified geological map of the MacVicar Lake–Ralph Anderson Lake area.

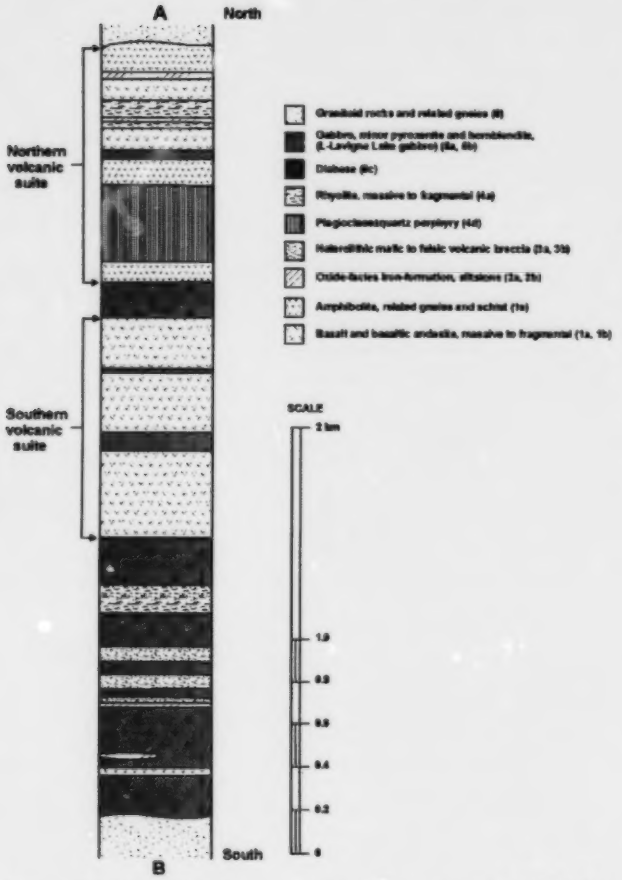

Figure GS-23-3: Transverse cross-section, oriented NNE-SSW through the Ralph Anderson Lake belt (marked as A-B on Fig. GS-23-2).

skialithic enclaves of basalt (up to 20 m wide) within granitoid rocks and associated gneiss. Pillow structure is preserved at one locality, but primary features are generally absent.

Heterolithic Volcanic Breccia and Associated Tuff; Related Sedimentary Rocks (Unit 3)

A unique occurrence of intermediate volcanic breccia (subunit 3b) on the south shore of central Aswapiswanan Lake contains attenuated felsic fragments (up to 25 by 2 cm) within an intermediate, tuffaceous matrix. The unit, more than 8 m in thickness, is assumed to be affiliated with volcanic breccia of inferred debris-flow origin, which is common within the greenstone belt farther to the west (Gilbert, 1999b).

INTRUSIVE ROCKS

Gabbroic units that are intimately interlayered with mafic volcanic flows in the MacVicar Lake–Ralph Anderson Lake area (subunits 1c, 1d) have been identified as synvolcanic on the basis of their field setting and geochemical signature (*see* 'Geochemistry' section). Major gabbroic intrusions (unit 5) in the same area have been interpreted as postvolcanic due, in part, to their typically massive texture, in contrast to the highly attenuated state of incorporated supracrustal enclaves, even though the mafic intrusions are locally foliated and sheared at the margins (Gilbert, 1999a). An alternative interpretation is that the gabbro intrusions (unit 5) are penecontemporaneous with mafic volcanism; according to this view, tectonic strain due to early, regional deformation (D_1) within the mafic sills was confined to incompetent zones (i.e. tabular skialithic enclaves and gabbro margins). The main, nonfoliated

Table GS-23-1: Geological map units in the southeast Max Lake area.

INTRUSIVE ROCKS

7 Leucogabbro, diabase

6 Granitoid rocks and related gneisses: tonalite, granodiorite, granite; minor plagioclase porphyry, pegmatite, aplite; hybrid gneiss derived from units 1 and 6
- (6a) Granite, massive; related aplite and pegmatite
- (6b) Granodiorite and granite, massive to gneissoid; minor K-feldspar blastic granodiorite-granite; minor pegmatite
- (6c) Tonalite and granodiorite, gneissoid
- (6d) Hornblende quartz diorite to diorite
- (6e) Tonalite, plagioclase-phyric; minor felsitic *lits*
- (6f) Hybrid gneiss (derived from units 1 and 6)

5 Gabbro, minor pyroxenite and hornblendite (Lavigne Lake gabbro; McLeod Narrows gabbro); diabase
- (5a) Gabbro, mesocratic to melanocratic
- (5b) Pyroxenite, hornblendite
- (5c) Diabase
- (5d) Magnetiferous quartz diorite, diorite

VOLCANIC AND SEDIMENTARY ROCKS

4 Rhyolite, massive to fragmental; heterolithic breccia, minor related sedimentary rocks; plagioclase±quartz porphyry (subvolcanic sill)
- (4a) Rhyolite, massive to fragmental
- (4b) Heterolithic volcanic breccia and tuff
- (4c) Volcanic-derived conglomerate, feldspathic greywacke and siltstone
- (4d) Plagioclase±quartz porphyry

3 Heterolithic volcanic breccia and associated tuff; related sedimentary rocks
- (3a) Heterolithic volcanic breccia and tuff, mafic to felsic fragments
- (3b) Heterolithic volcanic breccia and tuff, felsic and minor intermediate fragments
- (3c) Volcanic-derived conglomerate, greywacke and siltstone

2 Sedimentary rocks; altered supracrustal rocks
- (2a) Oxide-facies iron-formation
- (2b) Siltstone, feldspathic greywacke, minor chert
- (2c) Altered garnetiferous supracrustal rocks

1 Basalt, related fragmental and intrusive rocks; derived laminated amphibolite, schist and gneiss
- (1a) Aphyric basalt; minor plagioclase-phyric basalt and related gabbro
- (1b) Basalt pillow-fragment breccia, flow-top breccia
- (1c) Gabbro, minor hornblendite
- (1d) Gabbro, megaphyric to glomeroporphyritic
- (1e) Amphibolite, related gneiss and schist
- (1f) Spherulitic pillowed basalt

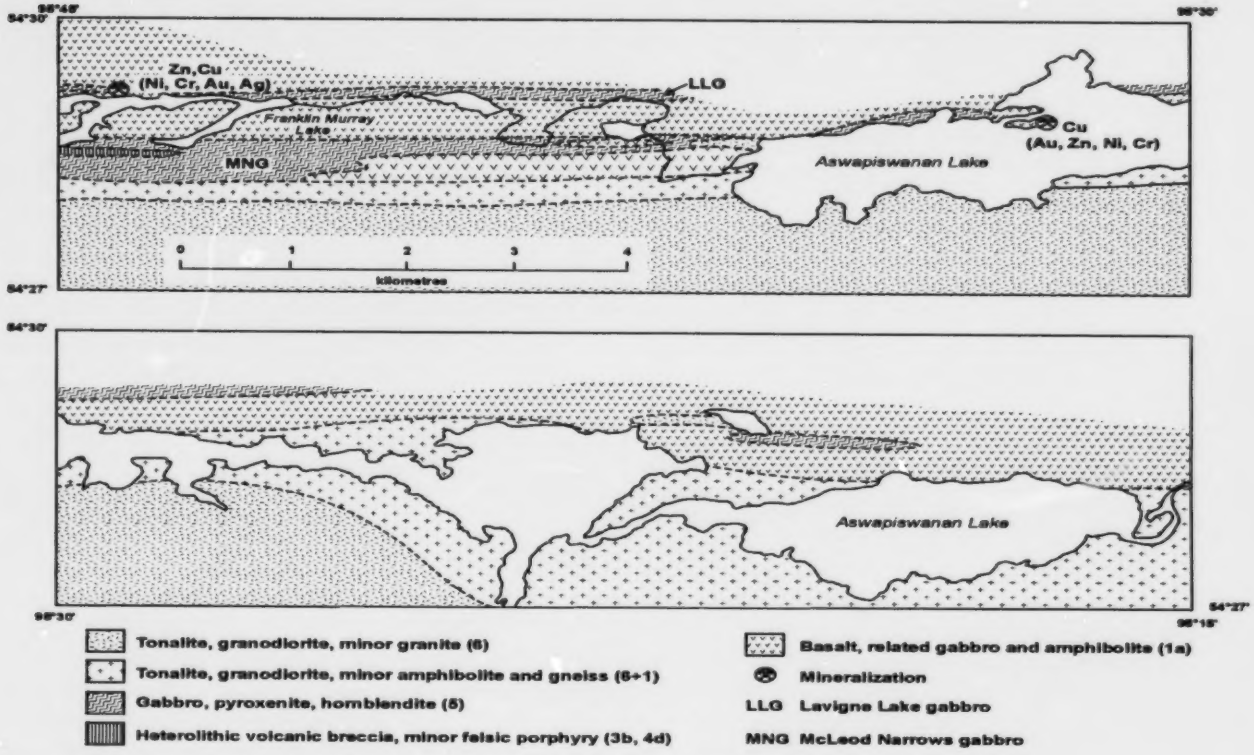

Figure GS-23-4: Simplified geological map of the Franklin Murray Lake–Aswapiswanan Lake area.

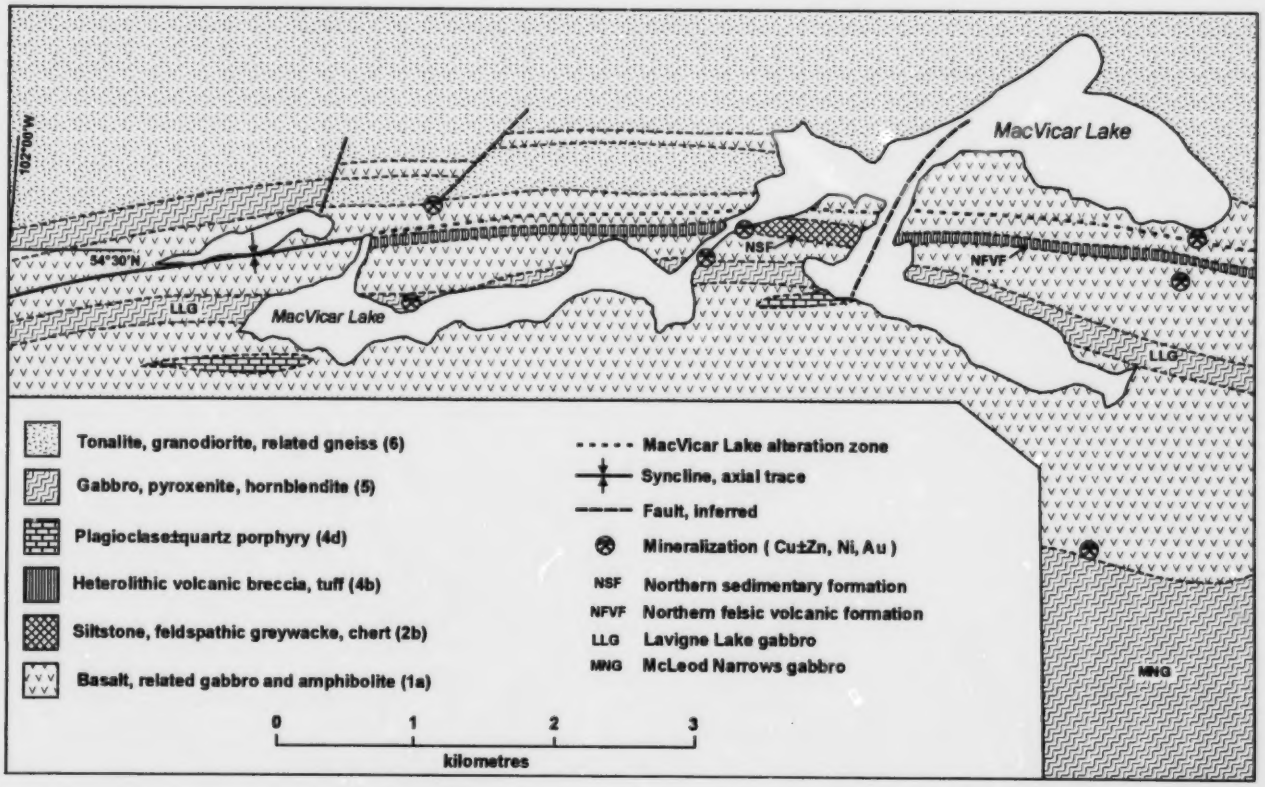

Figure GS-23-5: Map showing stratigraphic and structural details in the vicinity of MacVicar Lake in the northwest part of the project area.

cores of the mafic sills would thus be recrystallized but retain a massive fabric due to lack of strain within the central part of the intrusions. In support of this interpretation, the geochemical signatures of both the McLeod Narrows and Lavigne Lake gabbro intrusions (subunits 5a, 5b) are consistent with a magmatic affiliation between the gabbro and mafic volcanic rocks (see 'Geochemistry' section). Furthermore, basaltic enclaves (unit 1) within the gabbroic rocks (unit 5) are massive (i.e. there is no evidence that they differ significantly in age from the host rocks). In contrast, mafic volcanic inclusions within granitoid rocks (unit 6) are typically foliated and locally folded, indicating that a phase of metamorphism and deformation occurred prior to their incorporation. In summary, although the age of unit 5 gabbro intrusions cannot be clearly identified, there is some evidence that they are penecontemporaneous with mafic volcanism. Granitoid intrusions (unit 6) are invariably intrusive into gabbro (unit 5) and are therefore younger.

Gabbro, Minor Pyroxenite and Hornblendite; Diabase (Unit 5)

Massive, homogeneous gabbro (subunit 5a) in the Aswapiswanan Lake area occurs within the mafic volcanic section and as sporadic skialithic enclaves within younger granitoid rocks (unit 6) on the north shore of the lake. The mesocratic to melanocratic gabbro contains 40 to 70% (locally up to 90%) prismatic to equant hornblende (after pyroxene); leucocratic domains (<30% hornblende) are rare. One of the larger sills (>150 m thick) within the basaltic section on western Aswapiswanan Lake locally displays diffuse igneous layering (2–12 cm scale) defined by variable hornblende and plagioclase content. The gabbro is typically medium grained and massive to slightly foliated; sporadic minor shear zones, veined by quartz, are attributed to late brittle deformation.

Altered, olivine-bearing pyroxenite (subunit 5b), more than 120 m thick, occurs within mafic volcanic flows immediately north of the creek between Franklin Murray and Aswapiswanan lakes; the ultramafic rock extends laterally from the east end of Franklin Murray Lake for more than 1 km to the east. The unit is massive and homogeneous, except for minor brecciated zones (up to 1 by 5 m) that are characterized by ankeritic alteration. The brecciated zones appear to be the result of brittle fracturing, possibly penecontemporaneous with emplacement of the intrusion. Gabbro and hornblendite occur directly to the east, on strike with the pyroxenite. The ultramafic and gabbroic rocks at this locality are interpreted as part of the approximately 0.25 km thick Lavigne Lake intrusion (Gilbert, 1999a), which extends through the central part of the greenstone belt for a distance of more than 24 km, from MacVicar Lake in the west to central Aswapiswanan Lake in the east.

Granitoid and Related Gneissic Rocks: Tonalite, Granodiorite, Granite; Minor Plagioclase Porphyry, Pegmatite, Aplite; Hybrid Gneiss Derived from Units 1 and 6 (Unit 6)

Granitoid rocks of quartz dioritic to granitic composition extend along the south shore and part of the north shore of Aswapiswanan Lake, and occur as scattered minor intrusions within the predominantly mafic volcanic section that extends through the main part of the lake. The granitoid terrane south of the lake appears to be mainly granodioritic to granitic (subunit 6b). Along the south shore of Aswapiswanan Lake, a tonalitic to granodioritic zone, with localized gneiss and migmatite, extends between the mafic volcanic section to the north and the granodioritic-granitic terrane south of the lake. Within this 0.5 to 1.0 km wide zone, strongly foliated biotite tonalite with conspicuous subhedral plagioclase phenocrysts (up to 1 cm; 25% of the rock) extends intermittently along the entire south shore of the lake. The porphyritic tonalite (subunit 6e) occurs both as discrete concordant intrusions (approx. 1–10 m thick) and as a component of layered gneiss (subunit 6f) that contains diverse mafic to felsic *lits*. Gneissic layers (2–25 cm wide), which are interpreted as a product of the earliest tectonometamorphic event (F_1/M_1), are locally deformed by F_2 folds. The folds are truncated by younger granodiorite to granite (subunit 6b), which is massive to moderately foliated; this foliation, together with boudinage and disruption of granodioritic veins (subunit 6b), is attributed to a third (F_3) event. In southwestern Aswapiswanan Lake, the younger granodiorite is largely homogeneous, apart from localized white to pale pink, diffuse banding of inferred metasomatic origin. Porphyroblasts of K-feldspar locally occur in this unit.

The youngest granitoid phase is undeformed, and consists of massive granitic rocks (subunit 6a) that occur as minor dykes within earlier intrusions. The unit includes medium-grained granite, aplite and pegmatite; these occur both as exclusive rock types in some dykes and as contiguous, gradational phases in other intrusions. Minor mafic *schlieren* occur in the aplite-granite. Pegmatite is characterized by graphic quartz-feldspar intergrowths up to 25 cm across, randomly oriented biotite blades, and sporadic garnet. Rare traces of a surficial yellow bloom may be derived from a uranium mineral in the pegmatite.

Leucogabbro, Diabase (Unit 7)

Minor north-northeast-trending aphyric diabase dykes occur within gneissoid granodiorite (subunit 6b) on southwestern Aswapiswanan Lake. The mafic dykes truncate the granodiorite foliation, and contain an internal northeast-trending foliation that is attributed to local strain. These intrusions are interpreted as being related to north-northeast-trending leucogabbro-diabase dykes (unit 7), on the north shore of Ralph Anderson Lake, which represent the youngest intrusive unit in the map area (Gilbert, 1999a).

STRATIGRAPHY OF THE MACVICAR LAKE–RALPH ANDERSON LAKE AREA

Stratigraphic details of the MacVicar Lake–Ralph Anderson Lake area are the subject of a previous report on the geology of the southeast Max Lake area (Gilbert, 1999a). The following new information on selected stratigraphic units in the northwest part of the project area is complementary to that report.

Sedimentary Rocks: Feldspathic Greywacke, Siltstone; Oxide-Facies Iron-Formation; Altered Supracrustal Rocks (Unit 2)

A diverse assemblage of sedimentary rocks ('northern sedimentary formation') occurs within the northern mafic volcanic sequence in the northwest part of the map area, approximately 0.2 km south of the margin of the greenstone belt (Fig. GS-23-5). The sedimentary rocks wedge out to the west, where they are laterally gradational with felsic fragmental deposits (subunit 4b) of probable mass-flow origin. New structural data indicate that the sedimentary formation is located within a synclinal fold, the hinge line of which is probably coincident with a garnetiferous alteration zone (subunit 2c, Gilbert, 1999a) approximately 75 m north of the sedimentary rocks (*see* 'Structural Geology' section). The northern sedimentary formation is approximately 85 m wide on the peninsula in central MacVicar Lake, but the equivalent section farther west, at the limit of mapping, is only 40 m wide. To the east, the northern sedimentary formation is interpreted as laterally gradational with the northern felsic volcanic formation (Fig. GS-23-2; Gilbert, 1999a), which consists largely of massive to fragmental rhyolite with subordinate sedimentary units; the abrupt thickness change between these two formations at central MacVicar Lake is attributed to faulting.

Siltstone, feldspathic greywacke, minor chert (subunit 2b)

Fine-grained sedimentary rocks within the northern sedimentary formation consist largely of intermediate siltstone and greywacke with minor argillitic siltstone layers (subunit 2b). The rocks are both finely laminated (2–10 mm) and thinly bedded (2–40 cm scale), and are locally garnetiferous and/or characterized by stratabound cordierite and anthophyllite. Ovoid cordierite constitutes up to 65% of porphyroblastic units 10 to 50 cm thick; the blasts (up to 3 by 1 cm) are flattened within S_1 (Fig. GS-23-6) and locally lineated parallel to minor F_1 folds (Fig. GS-23-7). Randomly oriented, euhedral anthophyllite constitutes a later

Figure GS-23-6: Stratabound cordierite porphyroblasts parallel to S_1 in the northern sedimentary formation (unit 2).

Figure GS-23-7: Discordance between S_0 and S_1 in a cordierite-bearing siltstone-greywacke bed (subunit 2b) within the northern sedimentary formation. Cordierite is lineated parallel to F_1 folds.

metamorphic overprint (M_2). Sporadic argillitic, graphitic siltstone beds, commonly associated with amphibolitic interlayers, are associated with surficial iron staining due to pyrite oxidation.

Laminated chert is intercalated with greywacke and siltstone close to the south margin of the northern sedimentary formation, where chert beds up to 30 cm thick locally display flame structures at the contact with overlying greywacke (Fig. GS-23-8). Garnetiferous siltstone close to the north margin of the formation contains thin laminae of white chert and massive black magnetite (Fig. GS-23-9). Chert is also a subordinate component of garnetiferous mafic gneiss units (subunit 2c), which are located close to both north and south margins of the formation and typically display surficial iron staining.

Chert and related siliceous siltstone, up to 11 m thick, occur on the north shore of Franklin Murray Lake, in the upper part of the southern volcanic section. The south margin of the unit is mineralized where it is in contact with mafic volcanic flows (see 'Economic Geology' section); the north margin is in contact with a plagioclase porphyry sill.

A graded turbidite unit, at least 6 m thick, occurs within laminated amphibolite 60 m south of the contact between mafic volcanic and granitoid rocks in the northwest corner of the map area. Feldspathic greywacke and siltstone display graded A-D Bouma divisions, with well defined, south-facing tops (Fig. GS-23-10). The unit also contains stratabound cordierite, lineated parallel to F_1 minor fold axes.

Altered garnetiferous supracrustal rocks (subunit 2c)

A mineralized alteration zone, within the basaltic section in the vicinity of MacVicar Lake (Gilbert, 1999a), extends west to the limit of mapping in the northwest part of the map area; at that locality, the zone is less than 150 m from the north margin of the greenstone belt, at latitude 54.5°N (Gilbert, 2000). The lateral extent of the zone, together with stratigraphically equivalent oxide-facies iron-formation to the east, is more than 10 km; the thickness of the zone diminishes from a maximum of 36 m at the northeast corner of MacVicar Lake to 5 m at the west end of the lake. The alteration zone is coincident with a sedimentary unit that may overlie the mafic volcanic sequence in this part of the greenstone belt, based on the limited structural data available. The sedimentary unit consists of siltstone and chert (locally magnetiferous), intercalated with massive to laminated amphibolite. Whereas some amphibolite is interpreted as basalt, finely laminated hornblendic laminae within siltstone units are probably sedimentary in origin. Pervasive garnet blastesis, extensive silicic alteration, and surficial iron staining derived from pyritic zones are characteristic of both the siltstone and amphibolite. The alteration is interpreted to be the result of early hydrothermal activity and associated sulphide mineralization, probably contemporaneous with volcanism and related felsic porphyry intrusions. Porphyroblastic growth and subsequent disruption of the altered rocks due to sinistral strike-slip movement are attributed to later regional metamorphism and deformation. Lithological details of the MacVicar Lake alteration zone are given in Gilbert (1999a); assay data are discussed in the 'Economic Geology' section.

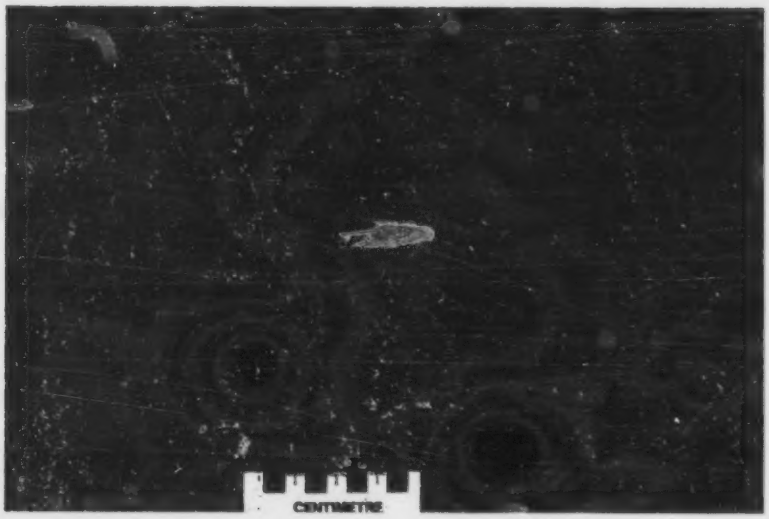

Figure GS-23-8: Flame structure in a chert layer (subunit 2b) within the northern sedimentary formation.

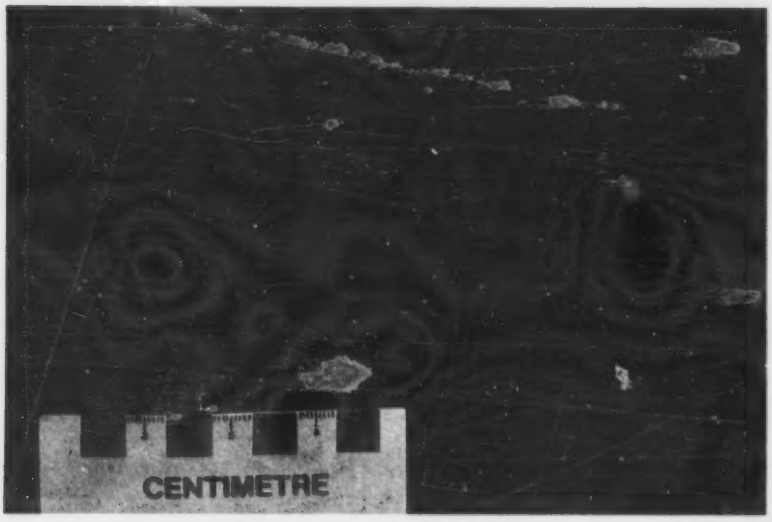

Figure GS-23-9: White chert and black magnetite laminae within intermediate to argillitic, garnetiferous siltstone (subunit 2b), close to the north margin of the northern sedimentary formation.

Figure GS-23-10: Graded turbidite deposit (subunit 2b) close to the margin of the Ralph Anderson Lake belt in the northwest part of the map area.

Layered garnetiferous gneissic zones (subunit 2c), approximately 1 to 4 m wide and similar to the MacVicar Lake alteration zone, occur close to both the north and south margins of the northern sedimentary formation. These zones are strongly deformed and characterized by pervasive metasomatism that resulted in silicic, hornblendic and chloritic alteration, extensive garnet blastesis, and development of massive garnetite. The garnetiferous zones, which display surficial iron staining due to the weathering of disseminated pyrite, appear to be mainly of sedimentary origin; amphibolitic layers may be derived, in part, from mafic extrusive or intrusive rocks.

Rhyolite, Massive to Fragmental; Heterolithic Breccia, Minor Related Sedimentary Rocks; Related Plagioclase±Quartz Porphyry (Unit 4)

A felsic fragmental rock formation, up to 34 m thick, occurs along strike from porphyroblastic siltstone (subunit 2b) of the northern sedimentary formation in the northwest part of the project area (Fig. GS-23-5). The felsic fragmental rocks are interpreted as laterally equivalent to porphyroblastic siltstone at central MacVicar Lake, and thus analogous to the northern felsic volcanic formation along strike to the east, north of Lavigne and Ralph Anderson lakes (Gilbert, 1999b; Fig. GS-23-2, -5). At outcrop scale, the felsic fragmental rocks directly overlie the porphyroblastic siltstone section at several localities in the central part of MacVicar Lake. The siltstone and related sedimentary rocks wedge out to the west, where they are replaced by the felsic fragmental rocks. The latter are interpreted as debris-flow deposits, possibly derived from pre-existing volcanic fragmental rocks akin to those in the northern felsic volcanic formation.

Heterolithic volcanic breccia and tuff (subunit 4b)

Heterolithic volcanic breccia in western MacVicar Lake is typically clast supported, with 40 to 75% diverse, felsic fragment types up to 1.2 by 0.3 m in size. The predominant white, rhyolitic clasts vary texturally due to differences in plagioclase phenocryst content (5–40% of clasts). A minority of clasts (<15%) are grey-weathering, intermediate types characterized by aphyric to sparsely plagioclase-phyric texture. The breccia formation is generally devoid of sorting, except for sporadic, diffuse, lapilli-bearing zones and felsic tuff interlayers (0.2–1.5 m thick). Breccia fragments are moderately to strongly flattened, and locally lineated with moderate east-southeast plunge. The shape of the original fragments is not known due to the deformation, but similar fragmental rocks in the northern felsic volcanic formation to the east display traces of original, angular clast shapes (Gilbert, 1999a). Fragmental rocks within the latter formation, which contain graded layers of possible turbidite origin, have been interpreted as subaqueous mass-flow deposits, and a similar mode of origin is inferred for the felsic fragmental rocks in western MacVicar Lake.

GEOCHEMISTRY

Fifty volcanic and related intrusive rocks in the west part of the Ralph Anderson Lake greenstone belt were analyzed for major, trace and rare-earth elements. Analysis of powder rock samples was carried out, using ICP–MS and ICP–OES, by Activation Laboratories, Ancaster, Ontario. The geochemical data are consistent with the field interpretation that the supracrustal assemblage is an arc-type volcanic sequence. A significant geochemical break divides the volcanic section into a lower, southern volcanic suite and upper northern volcanic suite, separated by the Lavigne Lake gabbro sill (Fig. GS-23-2, -3). The southern volcanic suite consists largely of basalt and synvolcanic gabbro that are depleted arc tholeiite. The southern basalt is characterized by LILE enrichment, high Th/Nb ratios and flat, incompatible element profiles with overall REE contents below those of depleted normal mid-ocean ridge basalt (N-MORB; Fig. GS-23-11A); in contrast, the more evolved basalt and

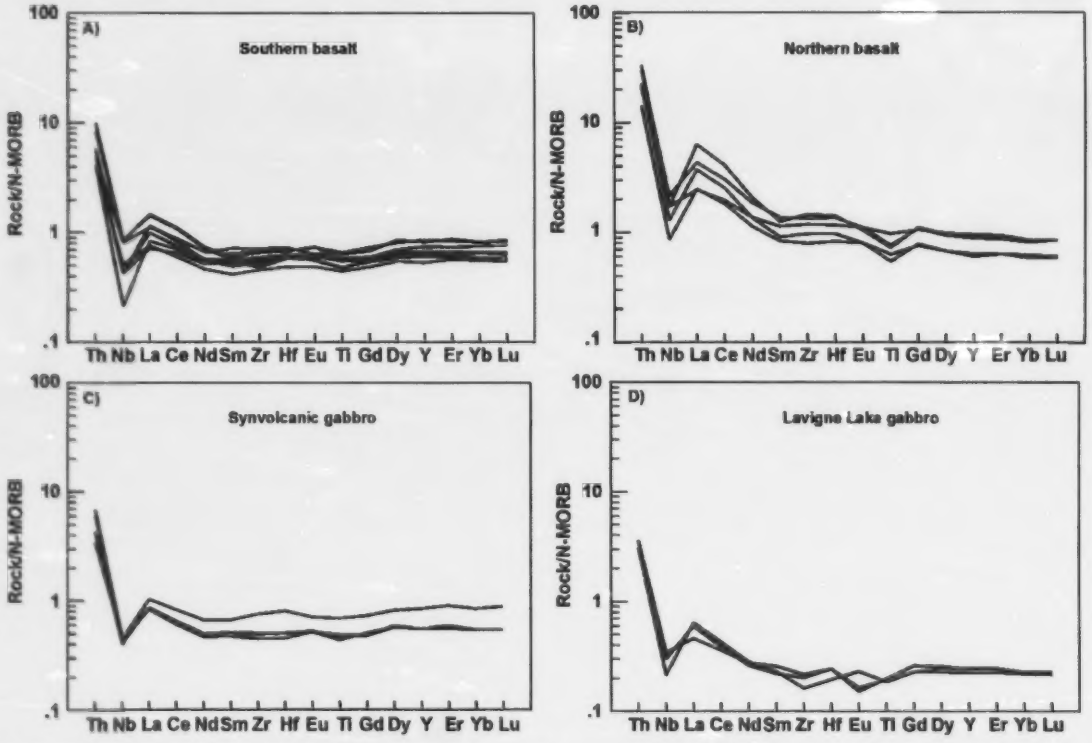

Figure GS-23-11: Extended element plots (N-MORB-normalized) of mafic volcanic and gabbroic rocks in the southeast Max Lake area, showing LREE (light rare-earth element)–depleted to –enriched, arc-type signatures of basalt and synvolcanic gabbro: A) southern basalt suite, B) northern basalt suite, C) synvolcanic gabbro, D) Lavigne Lake gabbro. Normalizing values from Sun and McDonough (1989).

basaltic andesite in the northern suite display moderately negative sloping, incompatible element profiles with more pronounced LILE/HFSE decoupling (Fig. GS-23-11B). These REE profiles and the FeO_{tot}/MgO vs. SiO_2 ratios (Fig. GS-23-12) indicate that the volcanic rocks are a transitional tholeiitic–calc-alkaline, arc-type assemblage. Both equigranular and megaphyric synvolcanic gabbroic units within the southern volcanic suite are compositionally very similar to the associated basalt; in contrast, the Lavigne Lake gabbro in the central part of the belt is relatively more primitive (Fig. GS-23-11C, D). The available geochemical data suggest that both the southern basaltic sequence and the Lavigne Lake gabbro are derived from a common, depleted, arc-tholeiite magma that was also the source for the northern volcanic suite.

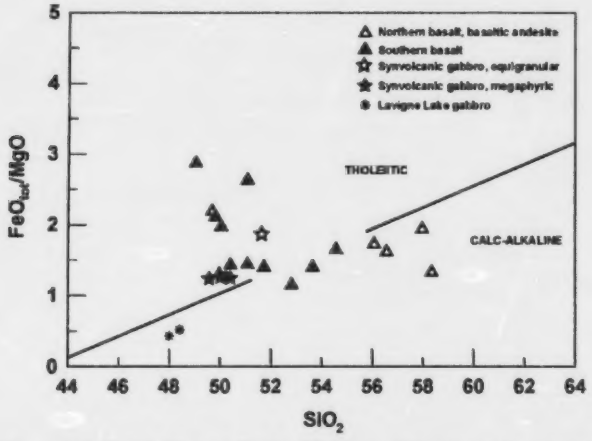

Figure GS-23-12: Plots of FeO_{tot}/MgO vs. SiO_2 for mafic volcanic and gabbroic rocks in the southeast Max Lake area.

Intermediate to felsic rocks in the west part of the Ralph Anderson Lake belt include massive to fragmental, volcanic extrusive types and plagioclase-quartz porphyry intrusions; geochemical data support the field interpretation that the two rock types are genetically related. Based on REE ratios and fractionation patterns, the volcanic extrusive and related intrusive rocks are classified as Archean 'F1 type' (Fig. GS-23-13), according to the scheme of Lesher et al. (1986); these rock types are characteristic of stratigraphic sequences that are devoid of base-metal sulphide mineralization, or barren horizons within mineralized sections (Lesher et al., 1986). Thus, the F1-type felsic volcanic extrusive rocks and most related porphyry dykes do not appear to represent good prospects for base-metal mineralization. Several silica-rich porphyry dykes at the north margin of the belt may be more prospective for economic mineralization, because some degree of crustal modification is indicated by SiO_2 contents that are elevated (>77%) above stoichiometric values. These intrusions are also characterized by strongly depleted TiO_2 (0.02–0.24%) and Nb (2–4 ppm), compared to other felsic porphyry and volcanic extrusive rocks in the map area (average TiO_2 = 0.5%; Nb = 7 ppm).

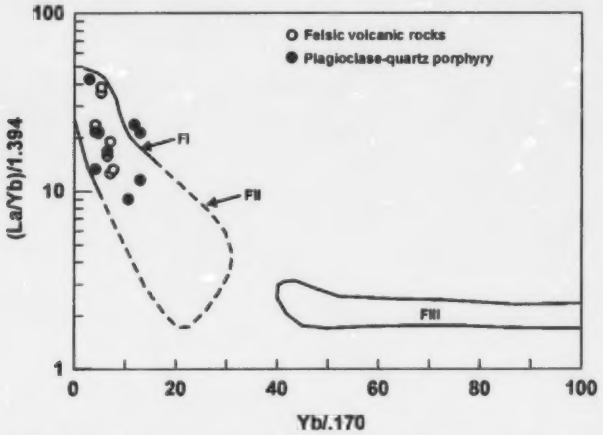

Figure GS-23-13: Plot of $(La/Yb)_N$ vs. Yb_N for felsic volcanic rocks and related plagioclase-quartz porphyry intrusions in the southeast Max Lake area. Fields of Archean felsic volcanic rock types after Lesher et al. (1986). Normalizing values from Sun and McDonough (1989).

STRUCTURAL GEOLOGY AND METAMORPHIC HISTORY

The structural and metamorphic history of the west part of the project area has been previously described (Gilbert, 1999a). The following is a brief summary of that account:

1) Major F_1 folding resulted in the subvertical, east-southeast trend of the volcanosedimentary sequence. Associated F_1 minor folds trend east-southeast with moderate plunge. The deformation was accompanied by regional S_1 foliation and associated planar features, such as clast flattening in fragmental rocks and lamination in amphibolite derived from basaltic rocks. Associated M_1 metamorphism resulted in the alteration of pyroxene to amphibole in mafic volcanic and intrusive rocks, and the development in sedimentary and altered volcanic rocks of various porphyroblasts, including cordierite (lineated parallel to F_1 folds; Fig. GS-23-7), andalusite and garnet (locally in massive garnetite laminae).
2) F_2 folding, localized boudinage of F_1 planar elements and refolding of F_1 folds were associated with regional movement parallel to S_1 (Fig. GS-23-14, -15). The F_2 folds plunge moderately southeast to east-southeast, roughly coincident with minor F_1 folds. The F_2 folding was associated with development of S_2 foliation (sub parallel to S_1), M_2 anthophyllite and cummingtonite porphyroblasts, and rotation and crenulation of M_1 porphyroblasts.

Figure GS-23-14: F_2 folding of S_1 foliation and related lamination in strongly attenuated volcanic rock.

Figure GS-23-15: D_2 boudinage and disruption of S_1 in altered basalt with felsic interlayers.

3) Discrete F_3 shear zones occur sporadically throughout the greenstone belt (locally at incompetent horizons such as brecciated basalt flow contacts, alteration zones or thin sedimentary rock units). The deformation resulted in localized conformable tectonic breccia zones, and mylonitization at the contact between the McLeod Narrows gabbro and mafic volcanic rocks immediately to the north (Fig. GS-23-2). Folding of S_2 foliation is associated with development of S_3 strain-slip cleavage. Associated M_3 metamorphism resulted in greenschist-facies alteration of earlier porphyroblasts, chloritization in shear zones, and localized silicification, carbonatization and epidotization throughout the greenstone belt, especially conspicuous in a zone along the south margin of the southern volcanic section.
4) The F_4 event resulted in north-northeast- to northeast-trending, high-angle faults (displacement on order of tens of metres) that transect the greenstone belt.

Structural data in the MacVicar Lake–Ralph Anderson Lake area indicate that the belt is a north- to north-northeast-facing monocline (Gilbert, 1999a, b). In the Franklin Murray Lake–Aswapiswanan Lake area, rocks of the southern volcanic suite dip moderately to steeply south; F_2 linear elements trend mainly southeast to east-southeast with moderate to steep plunge. New data from 2000 mapping at the west end of MacVicar Lake has identified an east-trending synclinal fold close to the north margin of the greenstone belt. At that locality, a turbidite unit faces south, in opposition to the predominant northward-facing direction throughout in the belt. The available data indicate that the hinge-line of the fold is approximately coincident with the MacVicar Lake alteration zone. Delineation of this structural feature in the area east of MacVicar Lake was not possible due to intense deformation and attenuation along the north flank of the greenstone belt.

ECONOMIC GEOLOGY

Seven mineral occurrence settings (Table GS-23-2) have been identified in the map area:

i) Hydrothermal alteration zone at central MacVicar Lake and associated oxide-facies iron-formations (at or close to the top of the northern volcanic suite)
ii) Stratabound mineralization associated with chert-siltstone units a) in the upper part of the southern volcanic suite; b) at the basalt/gabbro contact at the base of the northern volcanic suite; and c) within the northern sedimentary formation
iii) Felsic porphyry dykes within mafic volcanic rocks
iv) Gabbro-related mineralization, commonly close to margins of intrusions or at gabbro-basalt contacts
v) Sulphide mineralization in mafic volcanic rocks, occurring as irregular zones within pillowed flows, interstitial to pillows or in related breccia units
vi) Along the contact between the McLeod Narrows gabbro and the southern volcanic suite; mineralization is associated variously with siltstone, altered basalt and/or protomylonite in the contact zone
vii) The north margin of the northern volcanic section, at the contact between attenuated basalt and sheared, hybrid gneiss derived from the granitoid terrane to the north.

Details of mineralized localities listed in Table GS-23-2 are as follows:

i) Volcanic rocks in the MacVicar Lake alteration zone contain up to 420 ppm Cu and minor Zn (108 ppm). The mineralized samples are from strongly deformed layers of garnetiferous, silicified basalt with massive garnetite stringers and disrupted quartz veins. Magnetiferous chert beds within and adjacent to the alteration zone contain up to 314 ppb Au and 252 ppm Zn.
ii) Stratabound base-metal sulphide mineralization and gold occur in an approximately 11 m thick cherty siltstone unit close to the top of the southern mafic volcanic suite at Franklin Murray Lake. The mineralization (1700 ppm Zn, 470 ppm Cu, 31 ppb Au and traces of Ag, Ni and Cr) occurs over a 1 m wide section at the south margin of the sedimentary unit, which occurs between pillowed basalt to the south and a plagioclase porphyry sill to the north. At the south margin of the northern volcanic suite, a 2 m laminated chert unit between basalt and the Lavigne Lake gabbro at Ralph Anderson Lake contains 828 ppm Cu and traces of Au and Ag. Chert layers within the northern sedimentary formation at MacVicar Lake contain 293 ppm Cu and 237 ppm Zn.
iii) Between 250 and 400 m south of the MacVicar Lake alteration zone, quartz-plagioclase porphyry dykes within strongly deformed basalt contain up to 478 ppm Zn, 159 ppm Cu and a trace of Au.
iv) At Aswapiswanan Lake, gold and base-metal mineralization occurs at the south margin of a 10 m thick synvolcanic gabbro sill in the upper part of the southern volcanic suite (1373 ppm Cu, 306 ppm Cr, 124 ppb Au, and traces of Ni and Zn). The mineralization occurs in a narrow (30 cm wide) silicified zone in the marginal part of the sill. In the west part of the project area, both the Lavigne Lake and McLeod Narrows gabbro intrusions contain sporadic pyritic zones with Cu (687 ppm) and Zn (206 ppm).
v) Pillowed basalt, 150 m north of the Lavigne Lake gabbro, contains pervasive mineralized zones within and between pillows (up to 248 ppm Cu).
vi) At the south margin of the greenstone belt, a tectonized anthophyllite-chlorite-sericite alteration zone with patchy silicification contains 482 ppm Cu and 859 ppm Ni.
vii) At the north margin of the belt, minor Cu, Zn, Ni and Cr mineralization occurs in a narrow zone at the basalt-granite contact.

Table GS-23-2. Selected geochemical data for mineralized rock samples in the Ralph Anderson Lake greenstone belt.

Sample #	Type	Cu (ppm)	Zn (ppm)	Au (ppb)	Ag (ppm)	Ni (ppm)	Cr (ppm)	Location of mineralized zones	Host rock
									Quartz feldspar vein with chloritic laminae
32-99-2019-2	(i)	110	108	6	0.3	6	-	MacVicar Lake alteration zone, eastern MacVicar Lake	Silicified, tectonic laminated basalt; garnet-bearing, with mafic remnants and quartz veins
32-99-2020-1	(i)	424	8	-5	0.4	33	-	MacVicar Lake alteration zone, eastern MacVicar Lake	Silicified, tectonic laminated basalt; garnet-bearing, with mafic remnants and quartz veins
32-99-2020-3	(i)	49	82	-5	-0.2	27	-	MacVicar Lake alteration zone, eastern MacVicar Lake	Silicified, tectonic laminated basalt; garnetiferous amphibolite
32-99-2020-6	(i)	21	10	-5	0.5	6	-	MacVicar Lake alteration zone, central MacVicar Lake	Magnetiferous chert associated with garnetiferous chert component of oxide-facies iron-formation
32-00-3326-2	(i)	41	145	9	-0.3	13	59	MacVicar Lake alteration zone, west MacVicar Lake	Magnetiferous chert component of oxide-facies iron-formation
32-00-3281-1	(i)	23	252	314	-0.3	11	12	20 m south of the alteration zone, west MacVicar Lake	Cherty siltstone with fine bedding lamination
32-00-3023-2	(ii)	169	95	-2	-0.3	74	55	North shore, Franklin Murray Lake	Cherty siltstone with fine bedding lamination
32-00-3023-3	(ii)	288	1700	29	1.4	50	48	North shore, Franklin Murray Lake	Cherty siltstone with fine bedding lamination
32-00-3023-3	(ii)	178	206	-2	6.0	119	67	North shore, Franklin Murray Lake	Cherty siltstone with fine bedding lamination
32-00-3026-3	(ii)	470	976	6	1.2	78	50	North shore, Franklin Murray Lake	Chert/cherty siltstone associated with porphyroblastic siltstone
32-00-3026-4	(ii)	211	572	31	0.4	37	103	North shore, Franklin Murray Lake	Thin (0.5 m) chert unit within pillowed basalt adjacent to northern sedimentary formation
32-00-3026-5	(ii)	66	237	9	0.4	53	99	Northern sedimentary formation, MacVicar Lake	Chert unit between basalt and gabbro at base of northern volcanic suite
32-00-3234-3	(ii)	293	115	3	-0.3	43	56	Central peninsula, MacVicar Lake	Quartz-plagioclase porphyry in laminated amphibolite
32-00-3235-2	(ii)	828	64	19	1.6	65	-	North margin of Lavigne Lake gabbro, Ralph Anderson Lake	Quartz-plagioclase porphyry in partly altered and sheared basalt
32-99-2321-4	(iii)	99	476	57	0.3	26	-	North margin of MacVicar Lake alteration zone	Gossan zone within synvolcanic gabbro
32-99-2024-1	(iii)	159	134	-5	0.6	14	-	275 m south of MacVicar Lake alteration zone	Gossan zone within synvolcanic gabbro
32-99-2120-2	(iv)	202	146	-2	0.3	-1	-5	400 m south of MacVicar Lake alteration zone	Gossan zone at margin of synvolcanic gabbro
32-00-3104-1	(iv)	237	112	-2	-0.3	69	81	North shore, western Aswapiswanan Lake	Gossan zone at margin of synvolcanic gabbro
32-00-3111-1	(iv)	301	145	3	-0.3	6	-5	Island, northwest part of Aswapiswanan Lake	Gossan zone in gabbro adjacent to supracrustal enclave
32-00-3142-6	(iv)	1373	76	124	0.5	59	306	Island, northwest part of Aswapiswanan Lake	Gossan zone in gabbro adjacent to supracrustal enclave
32-00-3143-1	(iv)	166	119	13	0.4	-1	-5	McLeod Narrows gabbro (western part)	Gossan zone in melanocratic gabbro
32-00-3260-1	(iv)	177	206	-2	-0.3	28	26	Lavigne Lake gabbro (western part)	Silicified, epidotized 12 m alteration zone in gabbro
32-00-3264-1	(iv)	687	5	9	-0.3	70	-5	McLeod Narrows gabbro (western part)	Silicified, epidotized 12 m alteration zone in gabbro
32-00-3316-1	(iv)	246	38	-5	0.4	12	-	150 m north of central Lavigne Lake gabbro	Aphyric, pillowed basalt with spidotic domains and minor silicification
32-99-2222-1	(v)	482	29	-5	0.6	859	-	South margin of southern basalt suite	Altered amygdaloidal basalt (quartz–anthophyllite–sericite–chlorite–epidote); locally protomylonitic or gneissic
32-99-2041-3	(vi)							North margin of northern basalt suite	0.5 m gossan zone at contact between basalt and granitoid hybrid gneiss
32-00-3291-5	(vii)	66	146	4	-0.3	107	187		

Refer to text for explanation of mineralization types

Exploration Prospects

The numerous sulphide showings and the varied settings in which mineralization occurs in the southeast Max Lake area suggest that the terrane represents a viable exploration prospect for both base metals and gold. Bimodal volcanism, synvolcanic felsic porphyry intrusions, exhalative sedimentary horizons, conformable alteration zones and stratabound mineralization, all present in the southeast Max Lake area, are key features in VMS-hosting terranes of all ages, such as Noranda and Timmins (Gibson et al., 1984), and Flin Flon (Syme and Bailes, 1993). The stratigraphy in the southeast Max Lake area is also similar to that of some gold camps, such as Red Lake, Ontario (Pirie et al., 1982).

The best prospects for base-metal or precious-metal exploration are considered to be the garnetiferous alteration zone (unit 2c) that extends through MacVicar Lake, together with related magnetiferous iron-formation (type i); and stratabound, sediment-hosted mineralization in the central to upper parts of the sequence at MacVicar and Franklin Murray lakes (type ii). Alteration and mineralization in the garnetiferous zone may be a result of hydrothermal activity associated with synvolcanic felsic porphyry intrusions.

Copper-nickel mineralization at the south margin of the Ralph Anderson Lake belt may be related to the emplacement of the contiguous McLeod Narrows gabbro; however, the secondary mineral assemblage associated with this mineralization (anthophyllite-quartz-sericite-chlorite) is more typical of hydrothermal processes. The Cu-Ni mineralization occurs very close to an inferred major fault that could have provided a conduit for the hydrothermal fluids assumed to be the source of sulphide mineralization and alteration.

Lithostratigraphic and geochemical data suggest the southeast Max Lake area offers good potential for economic mineralization. However, this suggestion must necessarily be tempered by consideration of the fact that the geochemical signature of felsic volcanic rocks in the map area ('type F1' of Lesher et al., 1986) indicates that they have low potential for economic mineralization.

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GS-24 KINEMATICS OF THE NORTH AND SOUTH PIPESTONE LAKE SHEAR ZONES IN THE CROSS LAKE GREENSTONE BELT, NORTHWESTERN SUPERIOR PROVINCE

by A.C. Parmenter[1], S. Lin[1] and M.T. Corkery

Parmenter, A.C., Lin, S. and Corkery, M.T. 2000: Kinematics of the north and south Pipestone Lake shear zones in the Cross Lake greenstone belt, northwestern Superior Province; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 150-154.

SUMMARY

The Pipestone Lake area of the Cross Lake greenstone belt is transected by the east-southeast-trending north and south Pipestone Lake shear zones (NPLSZ and SPLSZ, respectively). Both the NPLSZ and the SPLSZ underwent two major generations (G_1 and G_2) of shear deformation, under amphibolite- and greenschist-grade conditions, respectively. The G_1 deformation in the SPLSZ was associated with south-side-up movement, but the kinematics of G_1 in the NPLSZ are not yet clear. The G_2 deformation in both shear zones was associated with subhorizontal dextral shear. Based on tracing of marker horizons and variation in bedding-foliation relationships and younging directions, a macro-scale tight fold was defined in the Cross Lake Group. The observation that massive anorthosite of the Pipestone Lake Anorthosite Complex cuts tectonic foliation in the mafic volcanic rocks of the Pipestone Lake Group indicates that the latter is significantly older than the anorthosite and that the two units are not comagmatic, as previously interpreted.

INTRODUCTION

A main focus of this project is to establish the structural framework of the Pipestone Lake area in the Cross Lake greenstone belt, with emphasis on the kinematic history of the north and south Pipestone Lake shear zones (NPLSZ and SPLSZ, respectively), two major east-southeast-trending structures that dissect the southeast arm of the greenstone belt. These shear zones are potentially part of a large-scale, east-trending shear zone system that extends from Manitoba to Ontario. The kinematic history and structural relationships established here will help to develop a regional tectonic model for the northwestern Superior Province.

Fieldwork in 2000 focused on 1:20 000 scale and detailed mapping of the area around Pipestone Lake. Stratigraphy and geological history of the area have been previously outlined by Rousell (1965), Corkery (1983, 1985), Corkery et al. (1988), Corkery and Lenton (1989) and Corkery et al. (1992). The Pipestone Lake Anorthosite Complex has been studied in detail by Cameron (1984, 1985), Peck et al. (1994) and Jobin-Bevans et al. (1995). Breedveld (1988) undertook a reconnaissance tectonometamorphic pilot study of the Cross Lake greenstone belt.

GEOLOGICAL SETTING

The Cross Lake greenstone belt is flanked to the northwest by the high-grade metamorphic Pikwitonei and metaplutonic God's Lake domains, and to the south by the metaplutonic Molson Lake Domain. Two major structural trends dominate the area: one, to the northeast, extends the length of Cross Lake and the other, to the east-southeast, follows the Nelson River system from west of the Whiskey Jack channel past the southeast corner of Pipestone Lake. Both the SPLSZ and the NPLSZ follow the east-southeast structural trend. Supracrustal rocks in the greenstone belt have been assigned to three main stratigraphic entities, the Pipestone Lake Group, Gunpoint Group and Cross Lake Group (Fig. GS-24-1; Corkery et al., 1992).

The Pipestone Lake Group (ca. (?)2760 Ma) includes predominantly metavolcanic and minor metasedimentary rocks, mainly extensive, weakly to strongly deformed pillowed basalt, minor massive basaltic flows, and thin ultramafic packages. It outcrops on both the north and south shores of Pipestone Lake. These two volcanic packages may not be genetically related and are hereafter referred to as the north and south Pipestone Lake groups, respectively. The north Pipestone Lake Group exhibits well preserved primary volcanic features and is only affected by the NPLSZ along its southern margin. It is bounded to the north by the Cross Lake Batholith. The entire south Pipestone Lake Group has been affected by the SPLSZ. It is bounded to the south by the Whiskey Jack gneiss complex and intruded by the Pipestone Lake Anorthosite Complex (PLAC), a layered complex of melagabbro, leucogabbro and megacrystic anorthosite. The PLAC is host to a major titanium deposit. The PLAC occurs between the south Pipestone Lake Group and the south Pipestone Lake granodiorite along the south shore of Pipestone Lake. Continuing west of Pipestone Lake, it curves to the southwest and follows the northeast structural trend along the southern extension of the Whiskey Jack channel.

The Gunpoint Group (ca. 2744–2728 Ma) unconformably overlies the Pipestone Lake Group and consists of fining-upward successions of metasedimentary rocks with interbedded felsic volcaniclastic rocks. A thin sliver of the Gunpoint Group outcrops along the south shore of Pipestone Lake and is bounded on both sides by Pipestone Lake Group volcanic rocks.

The Cross Lake Group (ca. 2713–2695 Ma), a generally fining-upward fluvial sequence, unconformably overlies both the Pipestone Lake and Gunpoint groups. Shoshonitic basalt and shoshonite-derived sediments were deposited with sediments of the upper Cross Lake Group. The Cross Lake Group is bounded to the north and south by Pipestone Lake Group volcanic rocks. These boundaries are defined by the SPLSZ and the NPLSZ.

SOUTH PIPESTONE LAKE SHEAR ZONE (SPLSZ)

The SPLSZ trends east-southeast. Its width varies from approximately 850 m at the Whiskey Jack channel to less than 100 m in the southeast corner of Pipestone Lake (Fig. GS-24-1). It affects the entire south Pipestone Lake Group, the Gunpoint Group, the northern margin of the south Pipestone Lake granodiorite, the southern margin of the Cross Lake Group, and the eastern part of the Pipestone Lake Anorthosite Complex. Evidence for deformation at both greenschist- and amphibolite-grade has been recognized, and two major generations (G_1 and G_2) of shear-zone deformation have been distinguished.

The G_1 deformation is associated with major shear movement under amphibolite-grade conditions. It is centred on the south Pipestone Lake Group, extending to its northern and southern contacts with the Cross Lake Group and the Whiskey Jack gneiss complex, respectively. Very strong deformation occurred in the southern part of the shear zone where mylonite is well developed. Mylonitic foliations dip moderately to the north along the Nelson River between the Whiskey Jack channel and Pipestone Lake, and steeply north or south along the southern shore of Pipestone Lake. Lineations plunge steeply downdip, both within and outside the mylonite zone. Layers of moderately to strongly sheared and epidotized pillowed basalt parallel the contact between the Pipestone Lake Group and the southern granodiorite. They are traceable along the

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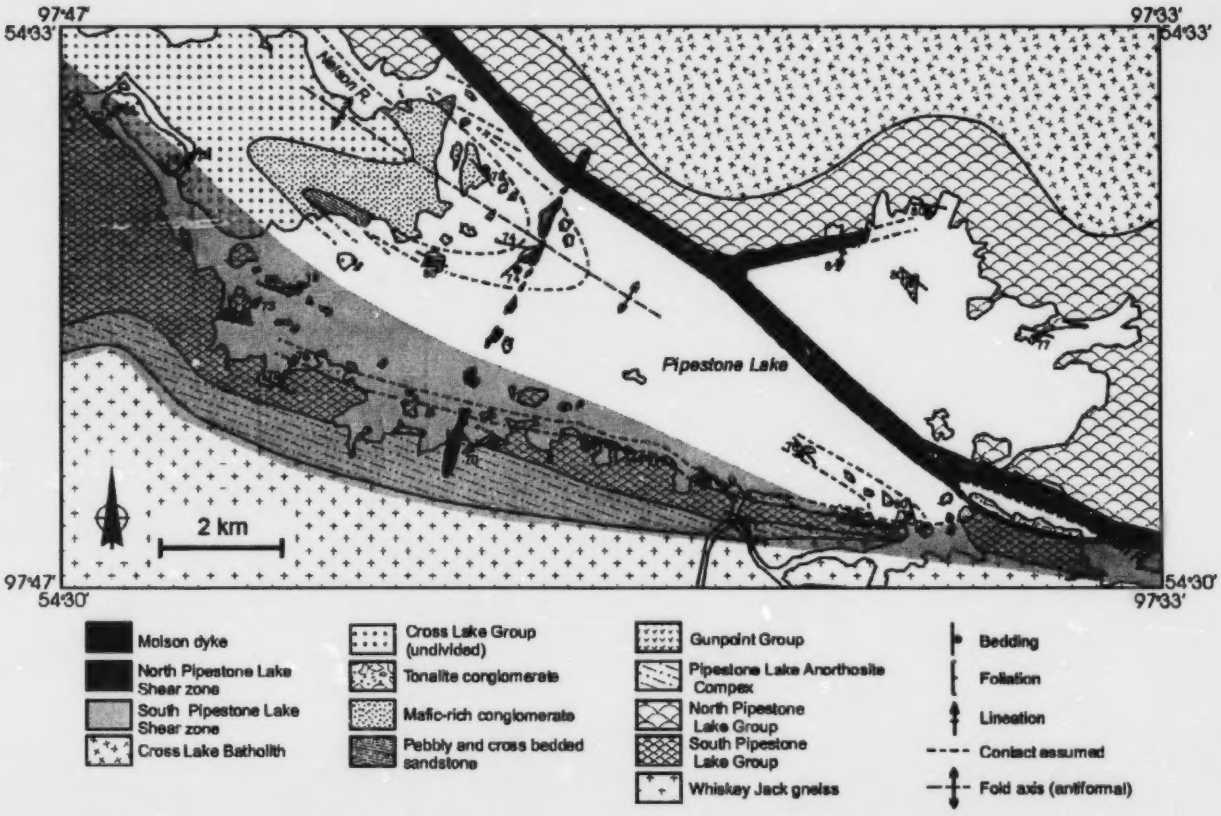

Figure GS-24-1: Simplified geological map of the Pipestone Lake area.

entire length of the SPLSZ and exhibit asymmetrical boudinage features. They indicate vertical south-side-up relative movement and dextral asymmetrical boudinage oblique to lineation (Fig. GS-24-2).

The G_2 deformation is associated with greenschist-grade, thin (<10 m wide) shear zones that overprint G_1 structures and are developed subparallel to the G_1 shear zone. The G_2 mylonitic foliations dip steeply north or south. Lineations are shallow to subhorizontal (east-plunging), approaching horizontal at the contact between the Pipestone Lake and Cross Lake groups in the southeast corner of Pipestone Lake. Abundant dextral shear-sense indicators are observed. These include asymmetrically boudinaged quartz veins in mylonitized basalt and S-C' structures in thin (<3m wide) chlorite-talc schist layers (Fig. GS-24-3).

In the southeast corner of Pipestone Lake, a layer of Cross Lake Group pebbly sandstone and conglomerate is sheared into the SPLSZ. Bedding strike in the layer varies from east-southeast to north, and younging direction changes from south-facing to west-facing toward the shear zone. Bedding-cleavage relationships indicate that this is related to an asymmetrical fold with its southern limb assimilating into the SPLSZ, consistent with dextral shear. At the shear contact, greenschist-grade mylonitization has obliterated any primary depositional features in the sedimentary rocks.

Quartz-carbonate mineralization is a characteristic feature of

Figure GS-24-2: Boudinaged epidotized pillow indicating south-over-north movement in the South Pipestone Lake Shear Zone (looking west).

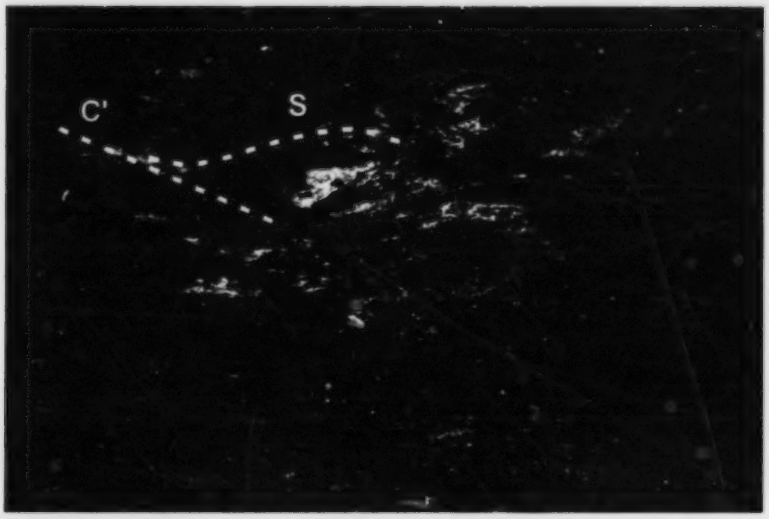

Figure GS-24-3: Chlorite-talc-schist in the South Pipestone Lake Shear Zone with S-C' structures indicating dextral movement.

greenschist-facies deformation in the SPLSZ. The quartz-carbonate occurs in mylonitized Pipestone Lake Group volcanic rocks, both within and adjacent to G_2 deformation zones. This may suggest a genetic link between carbonate-rich fluids and the G_2 shear event. Quartz-carbonate mineralization is most abundant through the centre of the SPLSZ, following the general distribution of the Gunpoint Group. Here, lenses of a distinctly black, carbonate-rich unit can be traced along the entire length of the SPLSZ. The distribution of this unit outlines macro-scale, steeply to shallowly plunging, s- and z-shaped, folded contacts between the Gunpoint and Pipestone Lake groups.

In the southeastern corner of Pipestone Lake, narrow (<40 cm wide), intermittent dextral shear zones deform a gneissic foliation in the southern granodiorite. Our preliminary interpretation links these minor shears with the G_2 deformation event because they have the same shear sense and the same metamorphic grade.

NORTH PIPESTONE LAKE SHEAR ZONE (NPLSZ)

The NPLSZ is developed in the north Pipestone Lake Group and exposed along the northern shore of Pipestone Lake and the southeastern extension of the Nelson River (Fig. GS-24-1). The exposed width of the shear zone varies from approximately 50 to 60 m. The true width of the NPLSZ is unknown because its southern boundary is not exposed. Coincident with the distribution of the NPLSZ are a thin (<4 m wide), discontinuous ultramafic layer and at least two generations of quartz-feldspar porphyry (QFP) dykes. As in the SPLSZ, evidence for both greenschist- and amphibolite-grade deformation has been recognized and two major generations (G_1 and G_2) of deformation have been distinguished.

The G_1 deformation is associated with vertical shear movement under amphibolite-grade conditions. Strong deformation produced a central 5 to 10 m wide mylonite zone. The mylonitic foliation trends east-southeast and dips steeply to the north or south along the north shore at the west end of Pipestone Lake. It curves toward the east-northeast and roughly parallels the north-central Pipestone Lake shoreline, where it is exposed for several hundred metres before being covered by water. Mineral lineations are steep in a downdip direction. Kinematics of the G_1 deformation are unclear due to the effects of the G_2 deformation.

The G_2 deformation is associated with a 20 to 30 m wide, greenschist-grade shear zone that trends east-southeast from northwestern Pipestone Lake to the Nelson River in the southeast. The G_2 mylonitic foliations dip steeply south and lineations plunge shallowly to subhorizontally toward the west. The S-C' structures in the ultramafic unit and boudinaged sedimentary clasts indicate dextral shear (Fig. GS-24-4).

Toward the southeast, the NPLSZ splits into two zones surrounding a thin wedge of conglomerate rich in tonalite clasts. This conglomerate only outcrops along the north shore of the southeastern extension of the Nelson River channel. It is very poorly sorted and dominantly clast

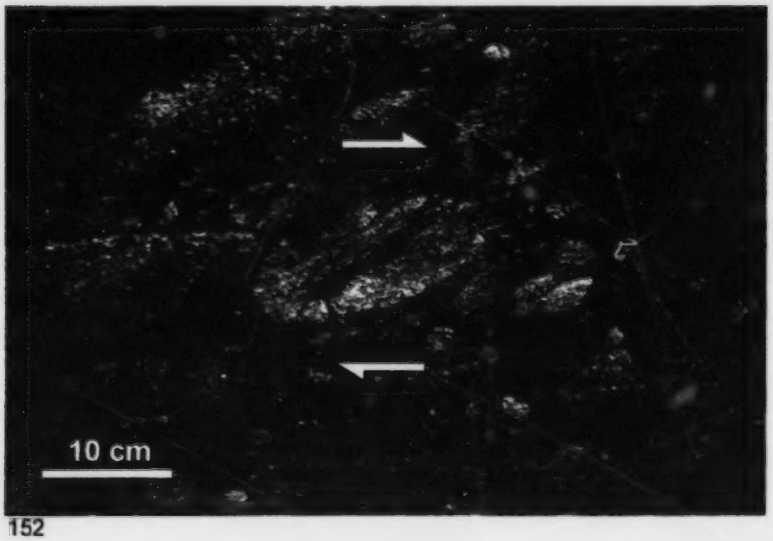

Figure GS-24-4: Dextrally sheared tonalite clast in the North Pipestone Lake Shear Zone, southeast Pipestone Lake.

supported, with approximately 80% tonalite clasts and 20% assorted granitoid and minor argillite clasts. Clasts range in shape from subrounded to angular and may exceed 1 m in diameter. The matrix is rich in medium- to coarse-grained subangular quartz with minor fine-grained mafic constituents. Toward the south contact with the NPLSZ, argillite is the dominant matrix constituent. This conglomerate is very different from any other sedimentary unit on Pipestone Lake. It most closely resembles a tonalitic conglomerate unconformably overlying the Town Tonalite (ca. 2719 Ma), approximately 20 km to the northwest. If these two conglomerates are correlative, it implies that a significant amount of movement occurred along the NPLSZ.

Primary volcanic features are well preserved in the Pipestone Lake Group in the northeast corner of Pipestone Lake. Flow-top indicators are abundant, indicating generally southward younging. Two distinct packages of volcanic strata can be traced with confidence over several kilometres (Fig. GS-24-1). They are defined by continuous layers of pillowed and massive basalt, amoeboid pillows, flow-top breccia, and characteristic plagioclase-phyric massive basalt. Dominantly prolate deformation occurred in these volcanic rocks, with localized L-tectonite development. The relationship between the structural development in this area and the kinematics of the NPLSZ will be a focus of further study.

STRUCTURES IN THE CROSS LAKE GROUP

Cross Lake Group metasedimentary rocks dominate west and central Pipestone Lake. While the north and south contacts of the group are sheared, most of the metasedimentary package contains well preserved primary sedimentary features. These features, including crossbedding and channel-fill structures, are excellent younging direction indicators and are useful, along with bedding-foliation relationships, in outlining a large-scale fold structure in the Cross Lake Group. It is a tight, steeply inclined, east-plunging fold with a variably developed east-southeast-trending axial-planar cleavage. The fold is best outlined by the contact between a clast-supported, mafic clast–rich conglomerate and an overlying pebbly sandstone unit (Fig. GS-24-1). The pebbly sandstone is gradational into a variably sorted, crossbedded sandstone that outcrops adjacent to a dyke belonging to the Molson Dyke Swarm in west-central Pipestone Lake, where the fold closure is located. On the north limb of the fold, the bedding strikes east-southeast and foliation is counter-clockwise to bedding. On the south limb, the bedding strikes east-southeast and foliation is clockwise to bedding. In the hinge area, the foliation is at a high angle to bedding. Bedding tops vary from north-younging on the north fold limb, to east-younging in the fold hinge and south-younging on the south fold limb.

Southward toward the SPLSZ, upper Cross Lake Group sedimentary rocks, including argillite-rich conglomerate, shoshonitic basalt, shoshonite-derived sedimentary rocks and turbidites, are generally south-facing but their relationship with this large-scale fold structure is still unclear. A meso-scale dextral drag-fold structure (Fig. GS-24-5) is developed in strongly foliated turbidites in the southwest corner of Pipestone Lake. The geometry is similar to that developed in sedimentary rocks at the southeastern contact with the SPLSZ. The relationship between deformation structures in the Cross Lake Group and the SPLSZ will be a focus of further study.

RELATIONSHIP BETWEEN THE PIPESTONE LAKE GROUP AND THE PIPESTONE LAKE ANORTHOSITE COMPLEX (PLAC)

The south Pipestone Lake Group and the PLAC have previously been interpreted to be comagmatic (Phinney et al., 1988; Corkery et al., 1992). This interpretation allows that the age of the PLAC (ca. 2760 Ma; Corkery et al., 1992) also represents the age of the south Pipestone Lake Group volcanism. Our detailed field observations do not support this interpretation. At two localities where the contact between the two units is exposed, massive megacrystic anorthosite cuts a strong tectonic foliation in a mafic unit of the south Pipestone Lake Group (Fig. GS-24-6). The observed geometry makes it impossible that the relationship is due to competency contrast. Elsewhere in the PLAC, similar megacrystic anorthosite has undergone variable amounts of deformation. This observation indicates that emplacement of the south Pipestone Lake Group volcanic rocks significantly pre-dated the PLAC and therefore must have taken place prior to 2760 Ma. It is possible that deformation in the south Pipestone Lake Group volcanic rocks at the contact with the PLAC, and elsewhere within the PLAC, are related to the G_1 deformation in the SPLSZ. If correct, this implies that the intrusion of the PLAC took place during G_1. These observations and interpretations should be considered in genetic modelling of the titanium deposit hosted by the PLAC.

ACKNOWLEDGMENTS

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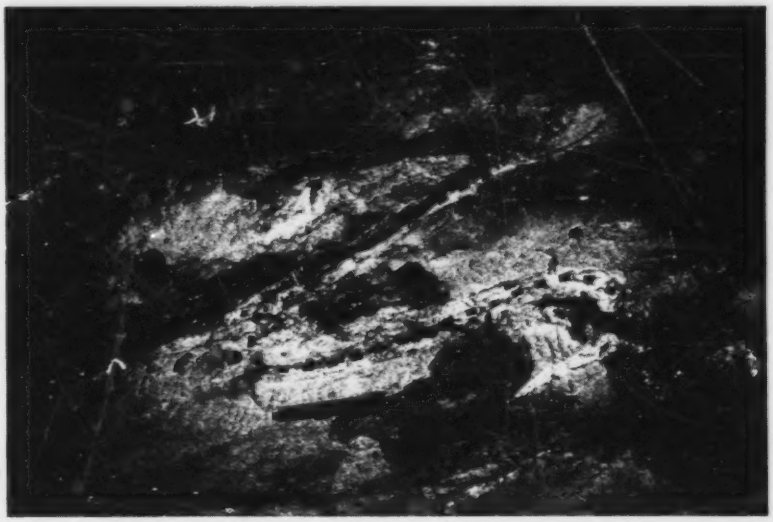

Figure GS-24-5: A Z-shaped drag-fold in Cross Lake Group turbidites, southwest Pipestone Lake.

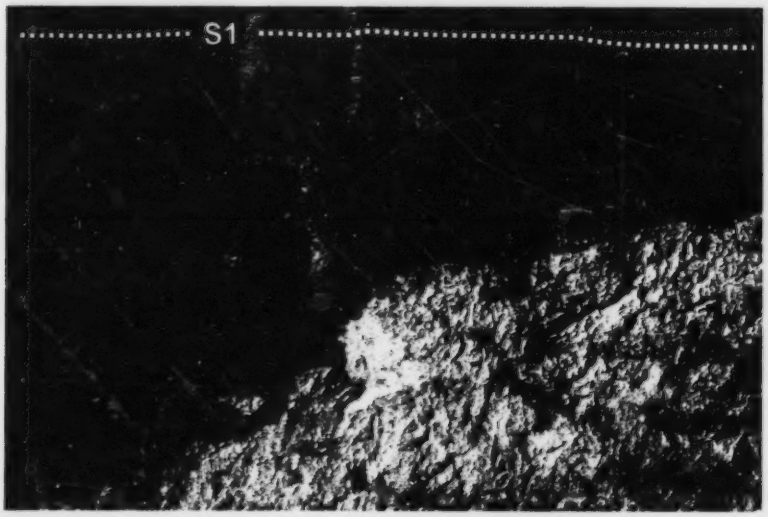

Figure GS-24-6: Massive megacrystic anorthosite crosscutting a tectonic fabric developed in a south Pipestone Lake Group mafic volcanic unit.

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ASSEAN LAKE ANCIENT CRUST: AN UPDATE

by M.T. Corkery, Ch.O. Böhm[1] and L.M Heaman[1]

Corkery, M.T., Böhm, Ch.O. and Heaman, L.M. 2000: Assean Lake ancient crust: an update; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 155-160.

SUMMARY

In 1997, the first hints of ancient crust were discovered at Assean Lake in northern Manitoba. This surprising discovery has led to extensive research in the Assean Lake area and the adjacent crustal domains of the Superior Boundary Zone. The Assean Lake ancient crust represents an Archean assembly of crustal segments that trend approximately 090 to 110° and are overprinted by Neoarchean and Paleoproterozoic deformation and metamorphism trending 060°. The Assean Lake crustal complex can be subdivided into 1) a southern, migmatitic, supracrustal series of quartz arenite, arkose and metagreywacke gneiss, containing amphibolite and silicate-facies iron-formation, and 2) a central, orthogneiss-dominated crustal segment that intrudes 3) a northern package of supracrustal rocks dominated by mafic to intermediate metavolcanic rocks and greywacke gneiss.

Combined Sm-Nd isotopic and U-Pb (by thermal-ionization mass spectometry [TIMS] and sensitive high-resolution ion microprobe [SHRIMP]) geochronological results indicate that Assean Lake preserves Paleoarchean to Mesoarchean crust that underwent a complex and prolonged history spanning more than two billion years.

INTRODUCTION

This program began in 1995 with the collection of samples in the Split Lake and Nelson River areas for U-Pb geochronology (Heaman and Corkery, 1996). It led to a detailed (1:50 000 scale) mapping project in 1997 at Assean Lake to better understand the Assean Lake deformation zone and the geological nature, kinematics and changes across this presumed Archean–Paleoproterozoic boundary (Böhm, 1997). Since 1997, Sm-Nd isotopic studies have been applied in conjunction with integrated mapping and reconnaissance programs on a regional scale. The Sm-Nd isotopic tool has proven to be ideal for discriminating between ancient (i.e. Paleoarchean to Mesoarchean), Neoarchean (i.e. typical Superior Province) and Paleoproterozoic (Trans-Hudson Orogen) crust. Based on the Nd isotope information, key samples were then analyzed by U-Pb dating methods in order to get primary and metamorphic ages. The combination of Nd isotope and U-Pb geochronological data has been instrumental in leading us to propose a new geological configuration of the craton margin, as indicated in Figure GS-25-1 and discussed in Böhm et al. (2000b). One of the most exciting results from this project so far is the identification of an assemblage of rocks in the Assean Lake area, previously mapped as Proterozoic, that is now known to be Mesoarchean (Böhm et al., 2000a). Evidence for ancient (pre-3.5 Ga) crust has provided enormous excitement for the possible preservation of some of the oldest rocks on Earth at Assean Lake. In addition, the re-interpretation of these rocks opens up potential for new mineral-exploration targets in this region.

The aim of this report is to focus on recent field work that provides an update on the component rock types, structural relationships and mineral potential of the ancient crust at Assean Lake. Together with

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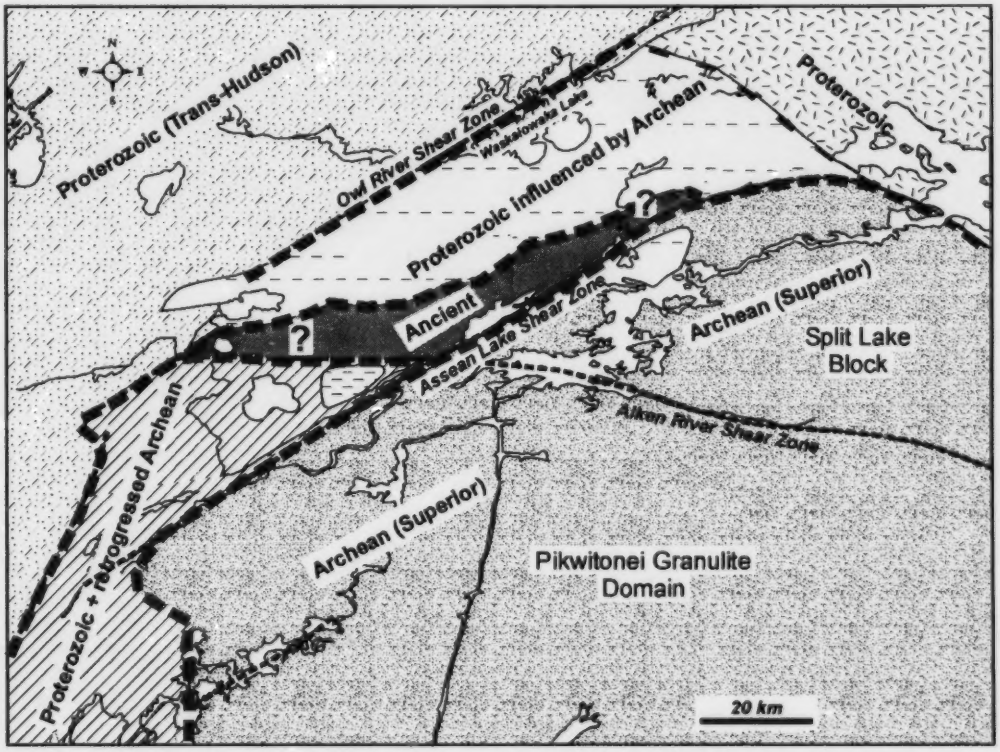


Figure GS-25-1: General geology and domain subdivisions in northwestern Superior Province.

geochemical, isotope and geochronological results, these new field observations should contribute to a better understanding of the nature and evolution of the Assean Lake ancient crust, and thus provide new insights into the early crustal evolution of the Earth.

GEOLOGY OF THE ASSEAN LAKE AREA

The more than 120 km^2 of ancient crust exposed at Assean Lake represents an Archean assembly of crustal segments that trend approximately 090 to 110° and are overprinted by Neoarchean and Paleoproterozoic deformation and metamorphism trending 060° (Fig. GS-25-2). The Assean Lake crustal complex can be subdivided into 1) the Clay River paragneiss, a southern, migmatitic, supracrustal series of quartz arenite, arkose and metagreywacke gneiss, containing amphibolite and silicate-facies iron-formation, and 2) a central orthogneiss-dominated crustal segment intruding 3) the Lindal Bay paragneiss, a northern package of supracrustal rocks dominated by mafic to intermediate metavolcanic rocks and greywacke gneiss.

Clay River Paragneiss and Related Intrusive Rocks

At the west end of Assean Lake, north of the main Assean Lake deformation zone, outcrop is dominated by paragneiss with subordinate orthogneiss (Fig. GS-25-2). Two colour varieties of paragneiss occur in this region (pink and grey). The high degree of recrystallization, mobilization and melt injection in the pink gneiss generally masks its probable sedimentary origin (i.e. derived from arkosic sedimentary rocks). Only the garnetiferous, greywacke-derived, grey gneiss is recognized as having a sedimentary origin. In addition, the paucity of outcrop and strong east-northeast-trending structural overprint associated with the Assean Lake Shear Zone preclude definition of mappable stratigraphy. Nevertheless, four distinct units can be described from this southern, migmatitic, supracrustal series.

The core of the Clay River paragneiss is a cordierite-sillimanite-garnet–bearing biotite metatexite (Fig. GS-25-3a). This migmatitic greywacke gneiss is interpreted to represent a sequence of thickly bedded, psammitic to semipelitic greywacke, most likely formed by density flows. Outcrops show 5 to 15% in situ mobilizate and minor associated garnetiferous pegmatitic injection. The Clay River metagreywacke is associated, on both the north and south sides, with a 5 to 10 m thick sequence of amphibolite. Decametre-scale layered amphibolite varies from massive, coarse–grained, salt-and-pepper–textured, hornblende-plagioclase amphibolite (Fig. GS-25-3b) to layered hornblende-diopside amphibolite interlayered with pelitic greywacke gneiss.

To the north and south of the amphibolite is a K-feldspar- and magnetite-bearing arkosic metatexite (Fig. GS-25-3c) that is strongly recrystallized and contains up to 25% mobilizate and injected granodioritic *lits*. This arkosic gneiss has previously been mapped as Sickle Formation (Lenton and Corkery, 1981). The compositional layering and interlayering with amphibolite of the Clay River paragneiss series are interpreted as being the result of primary sedimentary layering, migmatization and injection phases.

White-weathering, feldspathic biotite gneiss forms the southern and northern unit of the Clay River paragneiss (one outcrop displays a vague hint of trough cross-bedding). These gneissic units are highly variable in quartz content and probably represent quartz arenite composition in some layers (Fig. GS-25-3d).

Central Orthogneiss-Dominated Segment

The area north of the Clay River paragneiss series is dominated by a sequence of mainly 3.2 Ga tonalite to granodiorite intrusions and derived gneissic rocks. Similar to the Clay River paragneiss, the central orthogneiss underwent migmatization and was injected by later pegmatitic melt phases, resulting in variably layered orthogneiss (Fig.

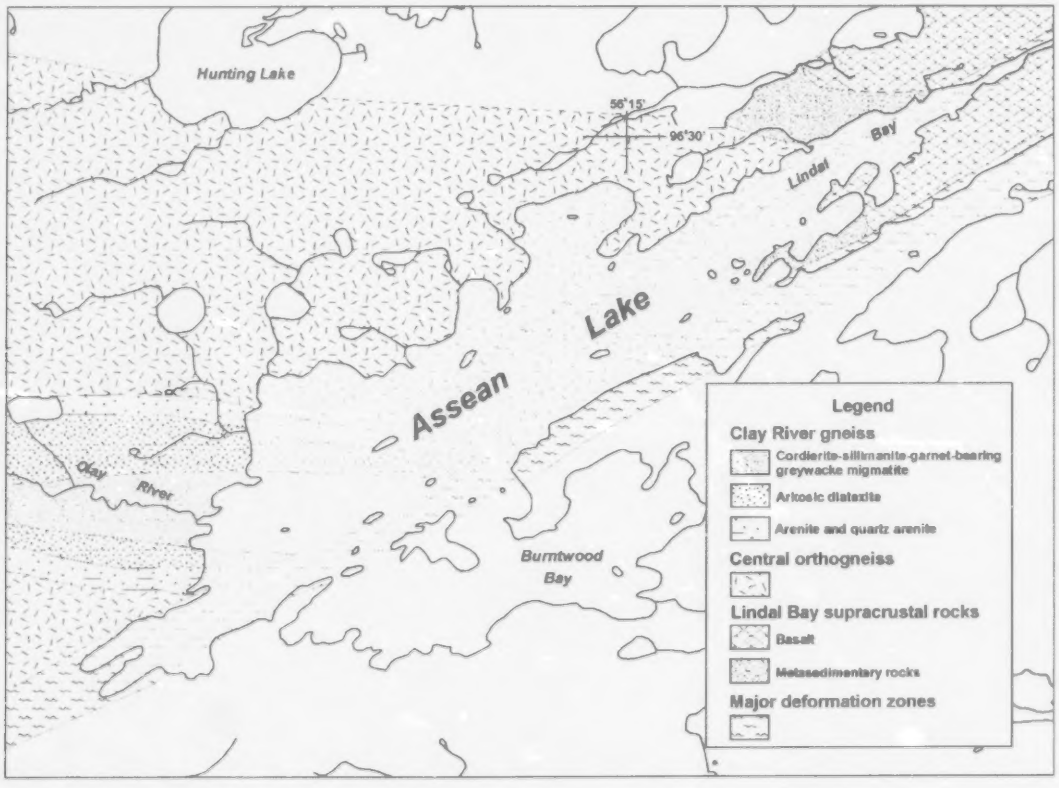


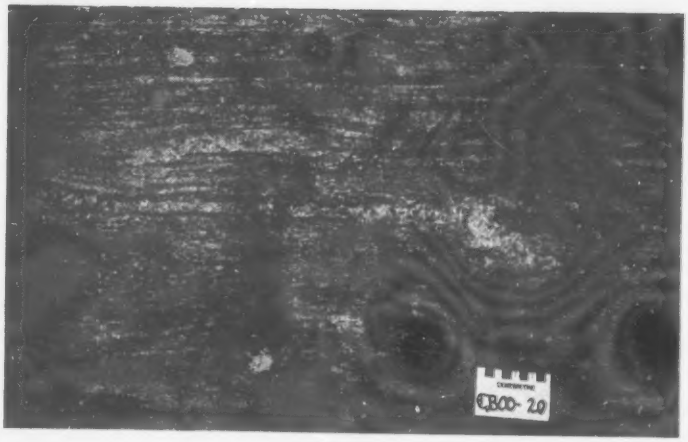
Figure GS-25-2: General geology of the Assean Lake area.

a)

Figure GS-25-3: Clay River metasedimentary rocks: a) cordierite-sillimanite-garnet–bearing greywacke migmatite, b) amphibolite, c) arkosic diatexite, d) arenite and quartz arenite.

b)

c)



d)

GS-25-4a, -4b). Larger, metre-scale, homogeneous pegmatite bodies seem to be directly related to pegmatite injection in the migmatitic orthogneiss. Relict lenses of paragneiss and minor interlayered amphibolite indicate that the central orthogneiss intruded an older supracrustal sequence. Centimetre- to decimetre-scale paragneiss lenses are generally fine-grained, garnet-bearing biotite gneiss. Detailed geochemical analysis of amphibolite samples is in progress and should address the question of whether these mafic lenses and layers represent relicts of early mafic dykes or form part of a supracrustal sequence that includes the above biotite gneiss.

Lindal Bay Supracrustal Sequence and Younger Intrusive Rocks

In the Lindal Bay area of northeastern Assean Lake, a package of supracrustal rocks dominated by mafic to intermediate metavolcanic rocks and subordinate metasedimentary rocks has been identified (Fig. GS-25-2). As in the Clay River area, a number of rock units can be defined, but stratigraphic relationships are difficult to determine due to the complex structural overprints and lack of continuous outcrop.

On the north shore of Lindal Bay, a sequence of rock units trending 090 to 110° is dominated by amphibolite, derived from mafic volcanic rocks, and associated gabbro sills with subordinate greywacke sedimentary rocks (Fig. GS-25-2). Near the contact with the central orthogneiss sequence to the southwest, granodiorite gneiss forms discrete layers in the supracrustal rocks, but contact relationships are overprinted. On the south shore of Lindal Bay, metasedimentary rocks dominate, with subordinate amphibolite, derived from mafic and ultramafic volcanic rocks, along the northwest margin. Rock units in this area are strongly re-oriented by the east-northeast-trending deformation zones.

The mafic volcanic rocks are generally highly deformed and vary from fine- to medium-grained, massive to centimetre-scale layered, hornblende-plagioclase amphibolite. In several locations, however, the layering can be recognized as remnants of pillow selvages (Fig. GS-25-5a). These rocks are predominantly basaltic in composition, with subordinate andesitic and ultramafic rocks. The ultramafic rocks occur as isolated outcrops with intrusive textures, and their age relationship to the paragneiss sequence is uncertain.

Metasedimentary units are generally thinly bedded and include quartz arenite, silicate-facies iron-formation, layered amphibolite, and mafic lithic and psammitic greywacke (Fig. GS-25-5b). Outcrops are generally small and most are dominated by one rock type. Stratigraphic relationships can only be observed in a few locations. At one location, quartz arenite grades into a garnetiferous plagioclase gneiss and then into thinly bedded, chert-anthophyllite-silicate iron-formation (Fig. GS-25-5c). The amphibole-rich layers become more abundant and the chert layers decrease, over a distance of 30 cm, into staurolite- and amphibole-bearing mafic pelite. This sequence then grades into thinly

a)

b)

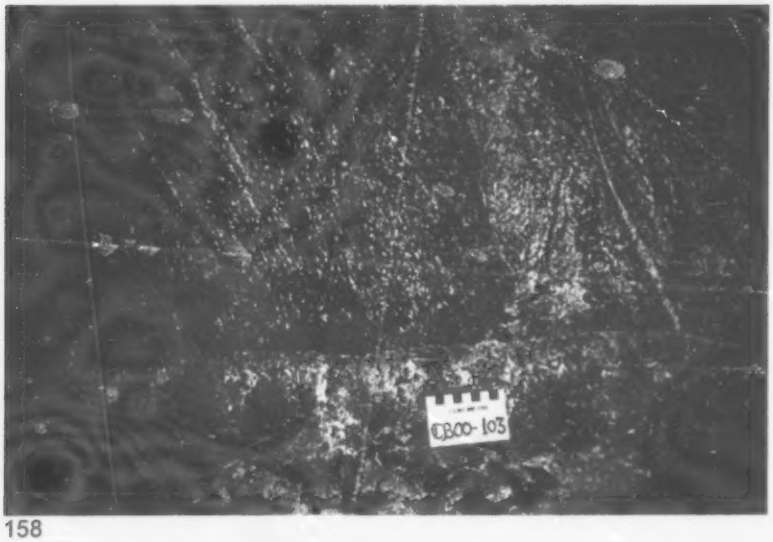


Figure GS-25-4: Central orthogneiss from the north shore of Assean Lake: a) hornblende-biotite granodiorite gneiss with pegmatitic melt injection, b) hornblende tonalite intruded by straight-walled pegmatite.

a)

Figure GS-25-5: Lindal Bay supracrustal rocks: a) highly strained pillowed basalt, b) staurolite-bearing pelitic metagreywacke, c) silicate-facies iron-formation.

b)

c)

layered staurolite- and biotite-bearing pelitic greywacke. Segments of this stratigraphy are observed on numerous outcrops on the peninsula between Lindal Bay and southeastern Assean Lake, but a sedimentary stratigraphy has not been determined. The contact relationships between the volcanic and sedimentary sequences are also unclear. However, both rock units seem to be intruded by later granodiorite to granite and pegmatite.

ECONOMIC CONSIDERATIONS

The Assean Lake area was found to have gold potential in the late 1930s (Lindal and Dunbrak veins), and a massive sulphide body was drilled in the 1960s. Lead has been known on Galena island since the early 1960s as well. Geological findings of this study may both explain and enhance the economic potential of the area. The Mesoarchean assemblage has the elements of a rift sequence, not unlike the Thompson Nickel Belt except in age. The recognition of zinc-lead showings in highly strained orthogneiss has until now been at most a curiosity. However, the recognition that the gneiss represents a paragneiss sequence may indicate that SEDEX-type deposits could occur in the area, as well as shear-hosted gold and nickel deposits.

The occurrence of thick ancient crust in this region may also provide a potential area for diamond-bearing kimberlite pipes.

DISCUSSION

Integrated mapping and geochemical, Sm-Nd isotopic and U-Pb geochronological studies across the northwestern Superior craton margin over the past five years have resulted in a re-interpretation of the location and nature of the boundary zone between Archean rocks of the Superior Province and Paleoproterozoic rocks of the Trans-Hudson Orogen in the region northeast of Thompson. Based on the discovery of ancient (pre-3.5 Ga) crust at Assean Lake, our research in this area has been intensified and continues to reveal surprising results. The mapping and sampling program during the 2000 field season was designed to document and better define the nature of the rock types that form part of the Assean Lake ancient crustal complex. The 2000 fieldwork identified three crustal assemblages as described above. Samples were collected from all major rock units for further geochemical, isotopic and geochronological studies. Highlights of the 2000 field program include the identification of two distinct supracrustal packages, separated by an orthogneiss-dominated central sequence. Within the supracrustal sequences, silicate-facies iron-formation and metabasalt flows provide critical stratigraphic horizons with which to trace the structurally complex and incomplete stratigraphic assemblies.

Unravelling the nature and history of a complex high-grade terrain such as the Assean Lake ancient crust requires a combination of detailed mapping, and isotope and geochronological tools. In addition to the continuous fruitful collaboration between researchers at the University of Alberta and the Manitoba Geological Survey, this project will benefit from collaboration with R. Stern, who is in charge of the sensitive high-resolution ion microprobe (SHRIMP) laboratory at the Geological Survey of Canada in Ottawa. For example, the first ion microprobe U-Pb age results from the Assean Lake ancient crust confirm the existence of crustal components as old as 3.9 Ga preserved in meta-greywacke detritus. Considering the size, complexity and heterogeneity of the Assean Lake ancient crust, this area clearly holds enormous potential for providing clues to the nature and evolution of Earth's earliest crust.

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GS-26 GEOLOGY AND STRUCTURE OF THE NORTH CARIBOU TERRANE-UCHI SUBPROVINCE BOUNDARY IN EASTERN MANITOBA, WITH EMPHASIS ON VOLCANIC AND VOLCANICLASTIC ROCKS OF THE BLACK ISLAND ASSEMBLAGE

by A.H. Bailes and J.A. Percival[1]

Bailes, A.H. and Percival, J.A. 2000: Geology and structure of the North Caribou Terrane-Uchi Subprovince boundary in eastern Manitoba, with emphasis on volcanic and volcaniclastic rocks of the Black Island assemblage; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 161-174.

SUMMARY

The Black Island assemblage of the Uchi Subprovince is exposed in a steeply dipping, north- and northeast-facing, greater than 3 km thick, monoclinal structural panel that records fabrics generated during juxtaposition with the adjacent, older North Caribou Terrane to the east. The lower third of the exposed Black Island assemblage consists of aphyric pillowed basalt and basaltic andesite, and overlying massive, porphyritic basalt and basaltic andesite (Gray Point sequence) that display normal mid-ocean ridge basalt (N-MORB) geochemical affinities. The upper two-thirds of the exposed Black Island assemblage consists of subaqueously deposited andesite and volcaniclastic rocks with calc-alkalic geochemical affinities (Drumming Point sequence). The abrupt change in tectonomagmatic setting from the N-MORB–type Gray Point basalt (?ocean floor–back arc) to calc-alkalic Drumming Point andesite flows and volcaniclastic rocks (?arc), combined with a basal Drumming Point volcaniclastic unit containing rounded fragments, suggests that the contact between the Gray Point and overlying Drumming Point sequence is an unconformity or disconformity.

The North Caribou Terrane, to which the Black Island assemblage is juxtaposed, comprises a ca. 3.0 Ga basement complex (tonalite with a layered igneous complex) and an unconformably overlying supracrustal sequence, the Lewis–Storey assemblage. The Lewis–Storey assemblage is composed of arkosic grit, quartzite, komatiite flows, iron-formation and carbonate. It is interpreted to represent a rift sequence similar to the less than 3.0 Ga and greater than 2.92 Ga Conley assemblage quartzite–carbonate–iron-formation–komatiite package, 80 km to the east at Wallace Lake (Sasseville and Tomlinson, 2000).

During juxtaposition of the Black Island assemblage and the North Caribou Terrane (at ca. 2.71–2.70 Ga), both the Black Island assemblage and the North Caribou Terrane were affected by D_1 structures and concurrent deposition of arkose and conglomerate (Hole River sedimentary rocks) in strike-slip basins during continued D_1 oblique convergence. Subsequent deformation was focused in shear zones during D_2 and D_3 transcurrent shear events.

Elucidation of the complex structural history along the boundary between the Uchi Subprovince and the North Caribou Terrane at the west end of the Rice Lake greenstone belt provides a much-improved geological context in which to conduct exploration activities. The previously unrecognized rift succession (Lewis–Storey assemblage) on the east shore and under Lake Winnpeg, in combination with the large areal extent of older ca. 3.0 Ga crust, opens up new exploration possibilities in this underexplored and previously poorly understood area.

INTRODUCTION

Mapping of excellent exposures on Black Island and along the adjacent east shore of Lake Winnipeg at 1:20 000 scale (Bailes and Percival, 2000; Percival and Bailes, 2000) was undertaken to characterize the nature of the boundary between two important tectonic elements of the western Superior Province (Fig. GS-26-1), the ca. 3 Ga continental crust of the North Caribou Terrane and the 2.99 to 2.70 Ga greenstone belts of the Uchi Subprovince. An understanding of the nature of this interface between the North Caribou Terrane and the younger greenstone belts of the Uchi Subprovince is important for tectonostratigraphic reconstruction of the western Superior Province and also to reveal controls on gold mineralization.

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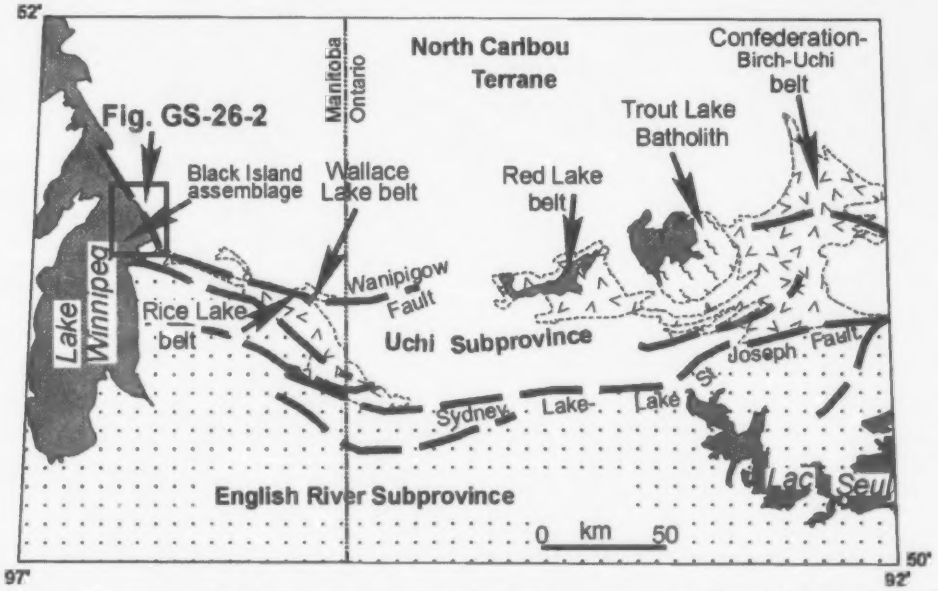


Figure GS-26-1: Major lithotectonic components of the western Superior Province in southeastern Manitoba and adjacent Ontario. The study area, shown by the rectangle, is located on the east shore of Lake Winnipeg and straddles the interface between the Uchi Subprovince and the older North Caribou Terrane.

This contribution documents the geology and geochemistry of volcanic rocks of the Black Island assemblage, part of the Uchi Subprovince. Other aspects of the geology, which are only briefly covered in this report, will be contained in subsequent contributions. Unit numbers used in this report correspond to those in Figures GS-26-2, -3 and -5 and on preliminary maps by Bailes and Percival (2000) and Percival and Bailes (2000).

GEOLOGICAL FRAMEWORK

Introduction

The Black Island assemblage, which outcrops in the southwestern part of the study area, forms the most westerly exposures of the Rice Lake greenstone belt in Manitoba (Fig. GS-26-1, -2). Together with the Red Lake greenstone belt of Ontario, these Uchi Subprovince volcanic rocks are important hosts for lode gold mineralization. Ermanovics and Wanless (1983) considered the Black Island assemblage to belong to the Neoarchean 2.73 Ga Bidou Lake subgroup, which is the dominant volcanic component of the Rice Lake greenstone belt.

The North Caribou Terrane, which forms the northeast and south portion of the study area, extends at least 400 km to the east (Fig. GS-26-1) and is one of the largest blocks with continental affinities in the western Superior Province (Thurston et al., 1991; Percival et al., 2000). The North Caribou Terrane is considered to be a Mesoarchean protocontinent (ca. 3.0 Ga; Thurston and Chivers, 1990) around which younger, Neoarchean supracrustal assemblages, including those of the Uchi Subprovince, assembled during accretion of the Superior Province (Stott and Corfu, 1991). More recent work by Rogers et al. (1999, 2000) and Sanborn-Barrie et al. (2000), however, indicates that at least some Mesoarchean–Neoarchean interfaces are conformable or unconformable contacts, rather than tectonic contacts.

North Caribou Terrane

East Shore Plutonic Complex

On the east shore of Lake Winnipeg, the North Caribou Terrane consists of a tonalitic basement (units 1–2, East Shore plutonic complex), unconformably overlain by a west-facing sedimentary-volcanic platform sequence (units 3–7, Lewis–Storey assemblage; Fig. GS-26-2, -3). The hornblende-biotite tonalite basement (unit 1) beneath the unconformity is coarse grained, homogeneous and characterized by blue quartz. It resembles tonalite and tonalite gneiss units dated in the vicinity at 3006 to 2998 Ma (Ermanovics, 1970, 1981; Krogh et al., 1974; Turek and Weber, 1994; V. McNicoll, unpublished data, 2000). The tonalite locally contains north-trending, chlorite-grade shear zones with prominent shear bands and, on the shore of Lake Winnipeg, grades to the east into a layered (on a 1–100 m scale) igneous complex (unit 2) consisting of sheets and pods of tonalite, quartz diorite, diorite and gabbro.

The ca. 3.0 Ga East Shore plutonic complex and English Lake complex share distinctive geochemical features. Both have rocks, ranging in composition from tonalite through diorite and gabbro, that have multi-element profiles with modest negative Ti and Nb anomalies and strongly depleted Th and Rb. These distinctive geochemical characteristics suggest generation from an ultradepleted mantle source.

The unconformable western margin of the East Shore plutonic complex is locally well preserved in zones of low strain on the shore of Lake

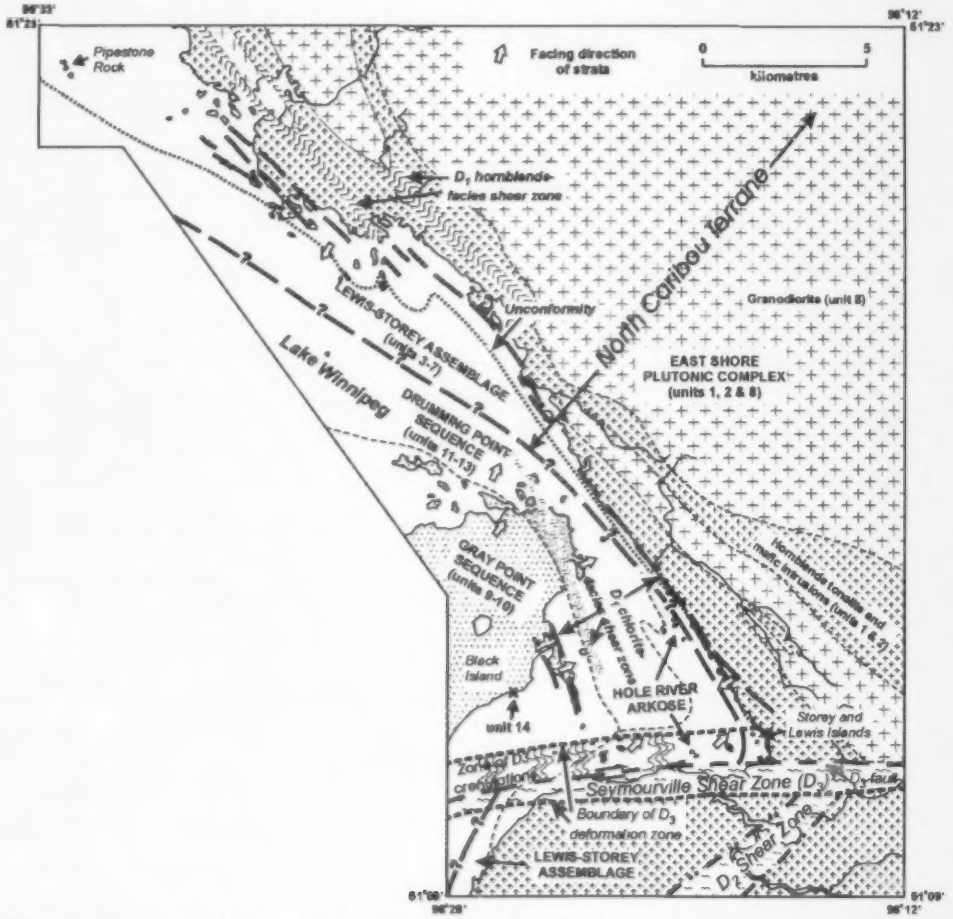


Figure GS-26-2: Simplified geology of the study area, showing the distribution of the main rock sequences and structures.

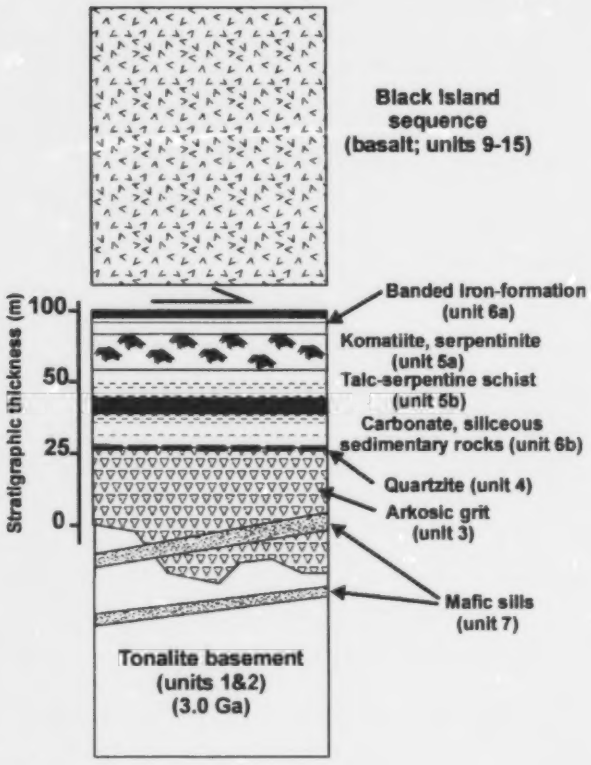


Figure GS-26-3: Schematic cross-section showing the stratigraphy of the Mesoarchean Lewis–Storey assemblage (units 3–7). The east-facing Lewis–Storey assemblage, a probable rift-platform sequence, lies unconformably on ca. 3.0 Ga tonalitic basement (units 1–2) and is in structural contact with Neoarchean volcanic rocks of the Black Island sequence (units 9–15). Unit numbers correspond to those on the preliminary maps by Bailes and Percival (2000) and Percival and Bailes (2000).

Winnipeg. Six exposures of the unconformity, on islands and shoreline outcrops on eastern Lake Winnipeg, allow it to be traced along strike for more than 30 km.

Lew–Storey assemblage

The East Shore plutonic complex is unconformably overlain, along its western margin, by a massive to thick-bedded grit (unit 3), consisting of angular quartz and plagioclase detritus with up to 10% chloritic matrix. Angular pebbles and cobbles of hornblende-bearing tonalite, identical to the basement, are locally present within a few metres of the base of this 15 to 40 m thick, generally unsorted grit unit (Fig. GS-26-4a). Thinly laminated, fine-grained quartzite (unit 4; Fig. GS-26-4b), which occurs in units up to a few metres thick in association with muscovite-rich schist, aluminosilicate-bearing schist and talc schist, is locally present as a thin veneer along the shoreline. Although a contact is not exposed, sedimentary rocks of units 3 and 4 are structurally overlain by komatiite (subunit 5a), which is exposed on Lewis Island. The komatiite displays well preserved spinifex textures defined by radiating clinopyroxene needles up to 20 cm long (Fig. GS-26-4c). Iron-formation (subunit 6a) on Storey Island, surrounded by serpentinite and talc schist (subunit 5b), consists of millimetre- to centimetre-scale layers of carbonate, chert, magnetite and sericite schist (Fig. GS-26-4d). Mafic sills (subunit 7b) up to 15 m thick intrude both the basement and cover sequence adjacent to the unconformity.

A prominent, positive, north-northwest-trending aeromagnetic anomaly that extends beneath Lake Winnipeg, from Storey and Lewis islands in the southeast to Pipestone Rock in the northwest, may delineate iron-formation (unit 6) and ultramafic rocks (unit 5), and the extension of the Lewis–Storey assemblage beneath the lake. The similar trend of the basal unconformity below the Lewis–Storey assemblage along the eastern shoreline of Lake Winnipeg is consistent with this interpretation of the magnetic anomaly.

The Lewis–Storey assemblage may correlate with the less than 3.0 to greater than 2.92 Ga Conley assemblage, a quartzite–carbonate–iron-formation–komatiite package 80 km to the east. This is supported by the presence of a similar, poorly preserved, sedimentary-volcanic sequence halfway between these two localities in the Wanipigow River area. At Wanipigow Lake, east-trending tonalite and supracrustal units resemble those observed in eastern Lake Winnipeg, although an unconformity was not noted. The tonalite is bordered to the south by a chloritic grit unit up to 10 m thick and is overlain by fine-grained, thinly laminated siliceous siltstone. Sporadic magnetite-chert and hematite-chert ironstone is also present at this structural level. Gabbroic sills with minor serpentinite schist layers occur structurally above the iron-formation, as well as within the tonalite. These units are variably highly strained, and are separated by the dextral Wanipigow Fault from a metagreywacke unit to the south (Weber, 1991).

The Lewis–Storey assemblage (units 3–7) is interpreted to be part of a platform-rift succession (cf. Thurston and Chivers, 1990) to the greater than 3.0 Ga protocontinent (East Shore plutonic complex, units 1–2). Basal grit (unit 3) may have developed through chemical weathering and slight downslope transport of basement units exposed as a result of plume-related uplift (cf. Rainbird and Ernst, in press). Deposition of quartzite (unit 4) and carbonate-bearing iron-formation (unit 6), accompanied by komatiitic magmatism (unit 5), probably reflects thermal subsidence during initial rifting. Younger mafic sills (subunit 7b) could be a distal expression of magmatism related to the drift phase of continental breakup.

Black Island Assemblage

The Black Island assemblage consists of lower greenschist facies volcanic and volcaniclastic rocks that form a steeply dipping, northeast-facing sequence, more than 2 km thick (Fig. GS-26-5), exposed in shoreline outcrops on Black Island and on smaller islands to the north (Bailes and Percival, 2000). The lower third of the section (units 9–10) consists of basalt and basaltic andesite flows (Gray Point sequence), and the upper two-thirds (units 11–13) consists of andesite flows and associated volcaniclastic rocks (Drumming Point sequence). Preliminary geochemistry (*see* 'Geochemistry of Volcanic Rocks, Black Island Assemblage' section) shows the Gray Point sequence to be MORB-like ([?] arc rift–ocean floor) and the Drumming Point sequence to be transitional calc-alkaline (?arc-like).

Ermanovics (1981) considered the Black Island assemblage (units 9–14) to correlate with the Bidou Lake subgroup exposed 65 km to the east. This appeared to be supported by a U-Pb zircon date of 2732 ± 10 Ma (Ermanovics and Wanless, 1983) from a rhyolite breccia (unit 14) on southeast Black Island and the ca. 2730 Ma age of the Bidou Lake subgroup (Poulsen et al., 1996) at Bissett. However, the rhyolite (unit 14) dated at Black Island has a faulted contact and an uncertain relationship to the main volcanic formations (units 9–13). The fault contact between the unit 14 rhyolite breccia and units 9 to 13 is significant, because rocks of unit 14 face south, whereas those of units 9 to 13 face north and northeast.

Hole River Sedimentary Sequence

The Hole River sedimentary sequence is composed of pink- and grey–weathering, sandstone–pebbly sandstone (unit 17) and rare conglomerate (unit 16). Uranium-lead zircon data on the Hole River sand-

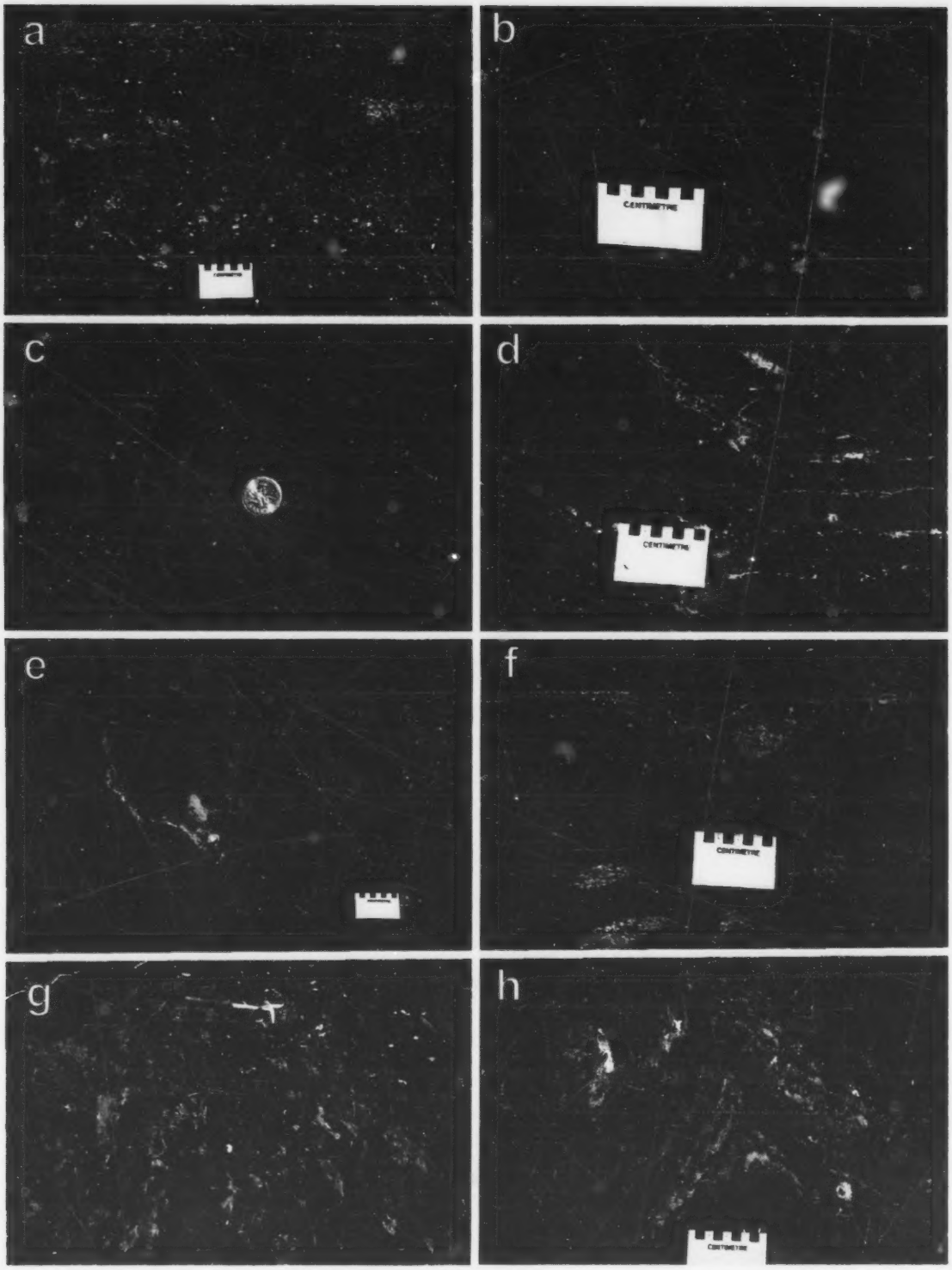


Figure GS-26-4: Photographs of the East Shore plutonic complex, Lewis–Storey assemblage, Hole River sedimentary rocks, and D_1 and D_3 structures: a) crude bedding in a pebbly grit arkose (unit 3) a few metres above the unconformity at the base of the Lewis–Storey assemblage; b) thinly laminated chert and quartzite (unit 4); c) komatiite (unit 5) from Lewis Island, with spinifex texture defined by radiating clinopyroxene needles; d) banded iron-formation (unit 6) from Storey Island, composed of thinly bedded magnetite (dark), carbonate and sericite schist; e) Hole River trough-crossbedded arkose (unit 17); f) Hole River conglomerate (unit 18) with prominent rounded granitic cobbles; g) D_1-shear zone developed in basalt on Gray Point; h) F_3-folds with prominent axial-planar crenulation cleavage deforming S_1-foliated basalt on island south of Gray Point.

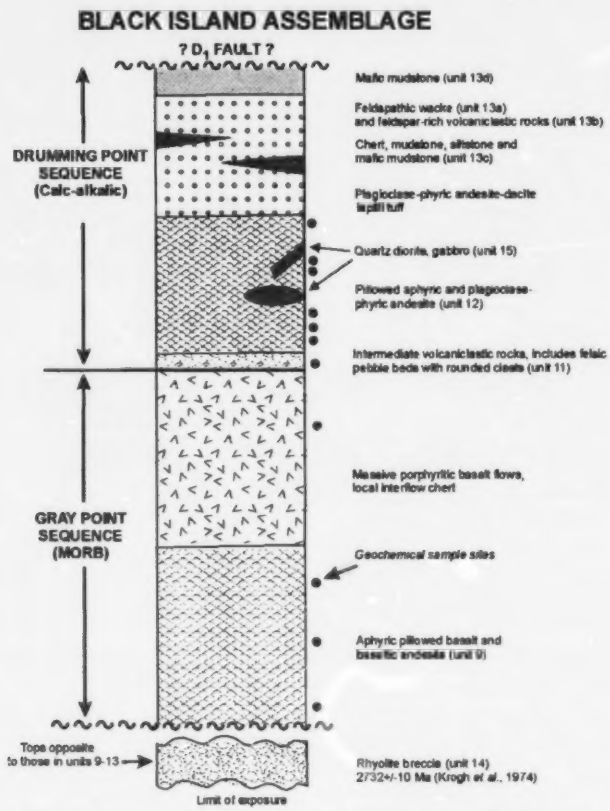

BLACK ISLAND ASSEMBLAGE
? D_1 FAULT ?
DRUMMING POINT SEQUENCE (Calc-alkalic)
GRAY POINT SEQUENCE (MORB)
Mafic mudstone (unit 13d)
Feldspathic wacke (unit 13a) and feldspar-rich volcaniclastic rocks (unit 13b)
Chert, mudstone, siltstone and mafic mudstone (unit 13c)
Plagioclase-phyric andesite-dacite lapilli tuff
Quartz diorite, gabbro (unit 15)
Pillowed aphyric and plagioclase-phyric andesite (unit 12)
Intermediate volcaniclastic rocks, includes felsic pebble beds with rounded clasts (unit 11)
Massive porphyritic basalt flows, local interflow chert
Geochemical sample sites
Aphyric pillowed basalt and basaltic andesite (unit 9)
Tops opposite to those in units 9-13
Rhyolite breccia (unit 14) 2732+/-10 Ma (Krogh et al., 1974)
Limit of exposure

Figure GS-26-5: Schematic stratigraphic section of the Black Island assemblage. Unit numbers correspond with those on preliminary maps by Bailes and Percival (2000) and Percival and Bailes (2000).

stone, obtained with the sensitive high-resolution ion microprobe (SHRIMP), show the youngest detrital grains to be 2706 Ma (V. McNicoll and J. Percival, unpublished data, 2000) and indicate a source and age similar to those of the Edmunds Lake and English River greywackes (<2704 Ma from detrital zircon geochronology; Poulsen et al., 1996) and the San Antonio formation sandstone (youngest concordant grain 2704 ± 14 Ma; Percival et al., 2000).

Hole River sandstone is typically massive, monotonous and indistinctly bedded, but locally displays excellent trough cross-bedding (Fig. GS-26-4e), pebble lags and scours. Pebbles in the sandstone are well rounded and include boulders and cobbles of a wide variety of plutonic rocks, vein quartz and volcanic rocks (Fig. GS-26-4f). Bed forms, rounding of pebbles and the wide variety of clasts in the Hole River sandstone suggest that these rocks were likely deposited in a subaerial-fluvial environment. Similar sandstone and conglomerate units of the San Antonio formation at Bissett, 65 km to the east, lie unconformably on volcanic (Bidou Lake subgroup) and plutonic rocks.

Structure

The Black Island assemblage, Lewis–Storey assemblage and adjacent portions of the East Shore plutonic complex carry a penetrative, north-trending, steeply west-dipping S_1 foliation and widespread concordant D_1 high-strain zones up to 40 m wide (Fig. GS-26-4g). These structures are interpreted to have formed in response to structural amalgamation of the Black Island assemblage with the older ca 3.0 Ga East Shore plutonic complex and its rift-margin Lewis–Storey assemblage. Shear-sense indicators, including extensional shear bands and asymmetrical boudinage structures, suggest consistent dextral kinematic sense on D_1 structures. Folds of S_1 foliation, previously considered a separate F_2 generation (Percival and Whalen, 2000), have consistent 'Z' asymmetry and can be shown in some high-strain zones to have formed during progressive development of the S_1 fabric; they are therefore termed F_{1a}. In the eastern part of the area, ductile D_1 fabrics are truncated by sheets and plutons of biotite granodiorite (unit 9) that may be correlative with ca. 2715 Ma granodiorite dated by Krogh et al. (1974). Southwest-trending crenulations (F_2) are common in thinly laminated high-strain (D_1) zones and are considered a distinct second generation of structures. A steeply dipping, east-trending zone of dextral greenschist-facies high-strain (D_3) and related crenulation folds (F_3; Fig. GS-26-4h) forms the linear shoreline in the vicinity of Seymourville. This structure translates the basement-cover sequence approximately 7 km to the west, where the basement hornblende tonalite (unit 2) is overlain by grit (unit 3) and komatiite (unit 5) in the Pelican Harbour area.

DESCRIPTION OF UNITS, BLACK ISLAND ASSEMBLAGE

The following description of volcanic and volcaniclastic rock units of the Black Island assemblage is largely derived from least deformed, unfoliated outcrops exposed on islands north of Black Island. The level of deformation of Black Island assemblage rocks increases toward the east and toward the contact with the adjacent North Caribou Terrane. More strongly deformed units on eastern Black Island and on islands to the east are characterized by a north- to north-northwest-trending S_1 foliation and local discrete D_1 shear zones. The boundary between the Black Island assemblage and the North Caribou Terrane is interpreted to be a D_1 shear zone, but lack of exposure prohibits verification of this interpretation.

Aphyric Basalt and Basaltic Andesite (Unit 9)

This unit, more than 1000 m thick, consists of aphyric basalt and basaltic andesite flows (subunit 9a) and mafic tectonite derived through D_1 strain (subunit 9b). The least deformed basalt and basaltic andesite of subunit 9a are exposed in small outcrops along the north and south shorelines of Black Island. These flows are characterized by medium to dark grey-green colour on both fresh and weathered surfaces, an absence of vesicles, and ubiquitous pillow structures with thin (<1 cm) selvages (dark green) and interpillow hyaloclastite domains (Fig. GS-26-6a). One pillowed flow displayed well developed thermal contraction cracks and radial pipe vesicles. Because of the small size of subunit 9a outcrops, no information is available on flow thicknesses or internal flow organization. No interflow sedimentary or tuff units were observed.

Basalt and basaltic andesite (subunit 9a) with S_1 foliation and associated north-trending and steeply dipping mafic tectonite and shear zones (subunit 9b) are exposed on Gray Point. Foliated pillowed basalt (subunit 9a) grades over a few metres into mafic tectonite (subunit 9b; Fig. GS-26-6b). On Gray Point, the D_1 mafic tectonite and shear zones contain numerous quartz-carbonate-chlorite vein stockworks (Fig. GS-26-6c) that were considered by Percival and Whalen (2000) to have potential for gold mineralization. These vein stockworks were sampled during the 2000 field season and submitted for gold analysis. Samples from two sites indicated Au values to be less than 8 ppb.

Porphyritic Basalt and Basaltic Andesite (Unit 10)

This 700 to 1200 m thick unit consists of massive basalt and basaltic andesite flows that weather pale buff-brown to rarely dark grey-green and are pale grey-green to rarely dark green on fresh surfaces. Least deformed exposures of this unit occur on islands directly north of Black Island; more deformed and less well exposed outcrops occur on small islands east of Gray Point. The flows lack amygdales and pillow structures, and contain only rare spherulites. Flow contacts are defined by chilled margins and by local interflow units of chert less than 3 m thick (Fig. GS-26-6d). Thicker massive flows commonly display subtle

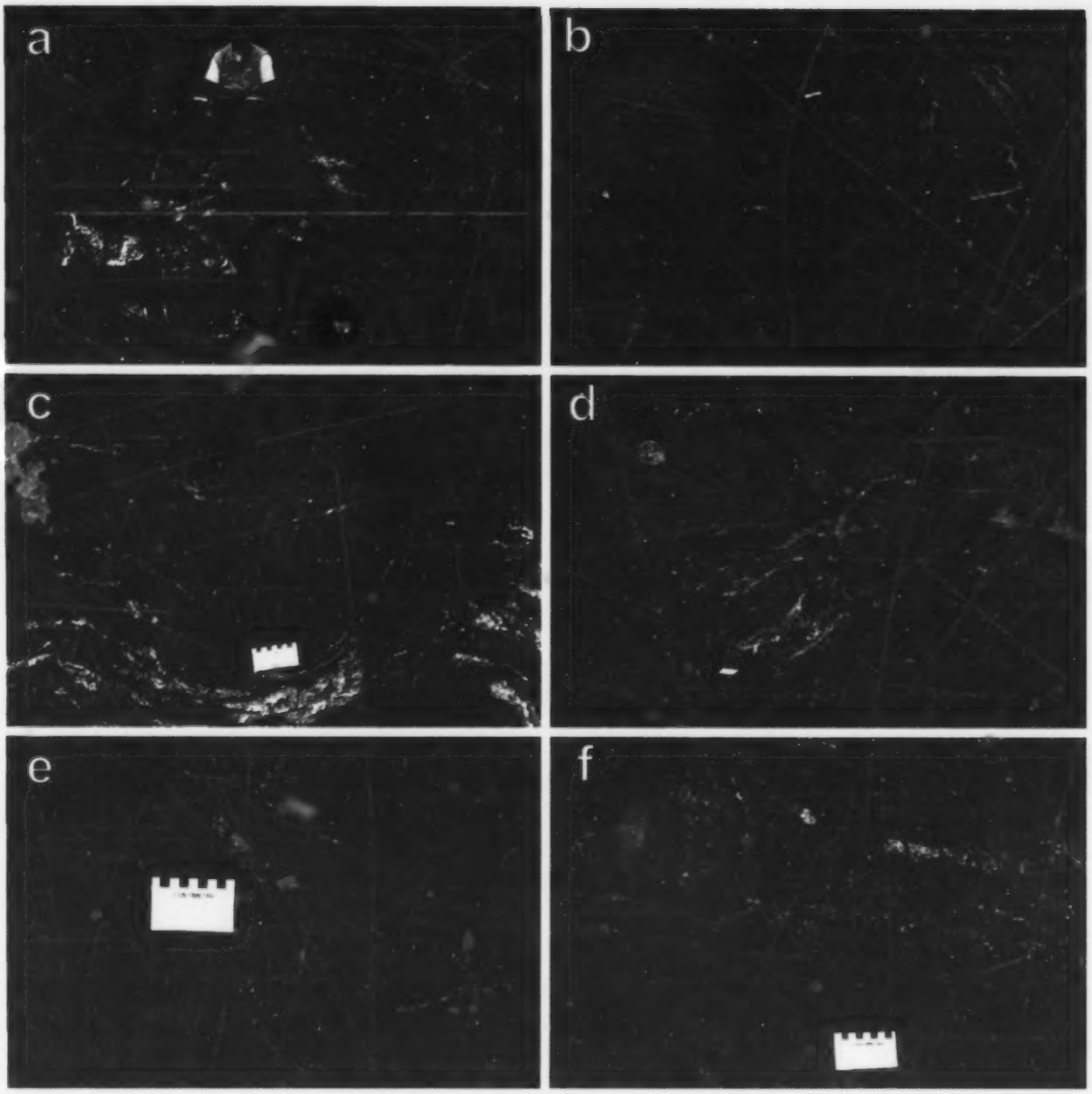


Figure GS-26-6: Photographs of rocks in the Point Gray sequence: a) pillowed basalt (unit 9) with D_1 flattening of pillows to left of person; b) contact between mafic tectonite and pillowed flow; c) quartz-carbonate vein stockworks in D_1 shear zone; d) interflow chert between two massive flows (unit 10); e) plagioclase megacrysts in massive basalt flow (unit 10); f) pegmatitic pod in massive gabbroic-textured flow (unit 10) prepared for collection of U-Pb radiometric dating sample.

variations in grain size that may be a consequence of incomplete cooling and amalgamation of flow units. Measured flows vary in thickness from 13 m to more than 60 m.

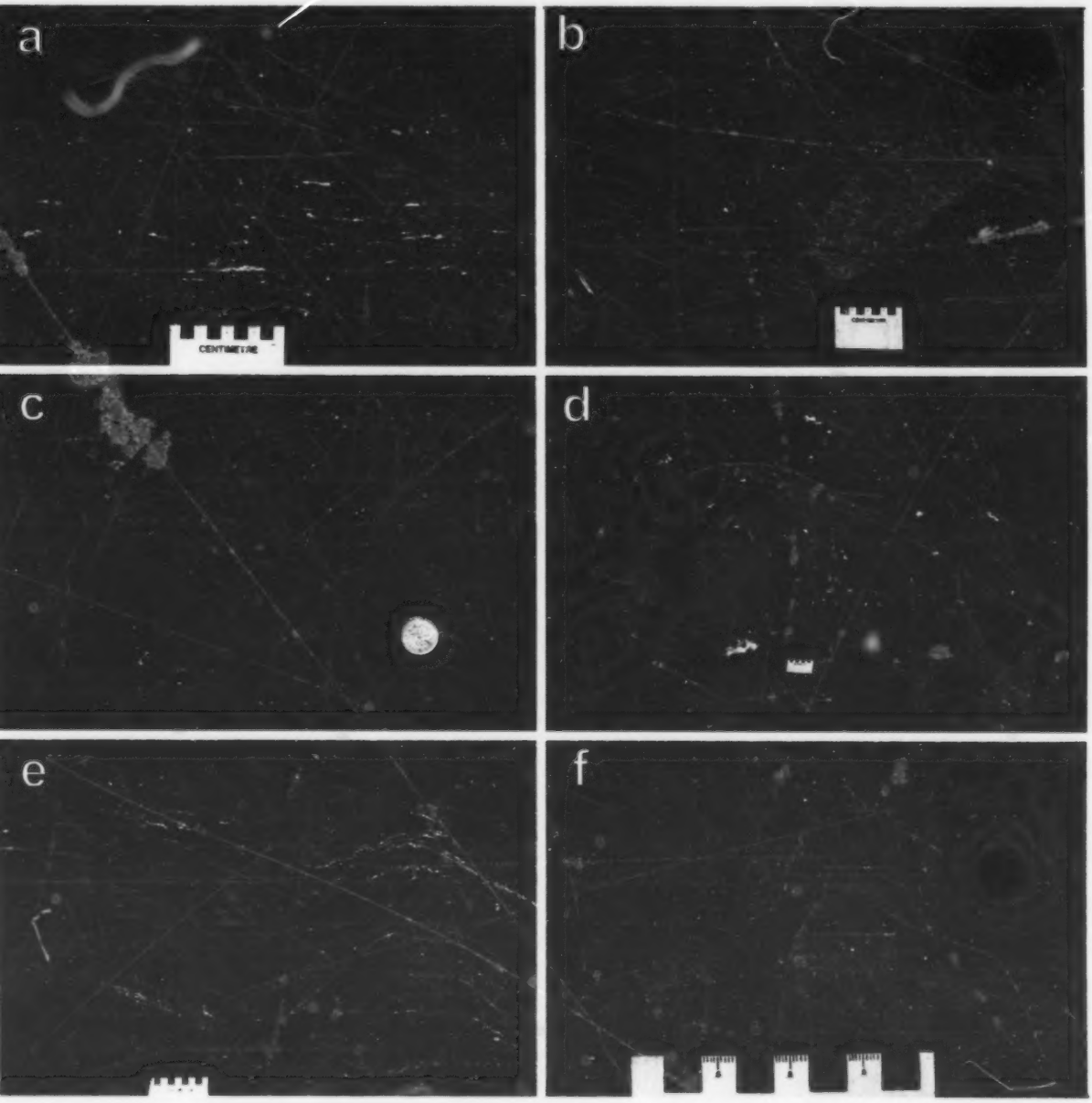
Most flows are pyroxene and pyroxene-plagioclase phyric, with up to 40% phenocrysts ranging from 2 to 8 mm in size (subunit 10a). Pyroxene phenocrysts in subunit 10a flows are often oikocrystic. Less common, but still abundant, are plagioclase megacrystic (subunit 10b) and pyroxene glomerocrystic (subunit 10c) flows. The plagioclase megacrysts that make up less than 7% of plagioclase-pyroxene–phyric flows of unit 10 are typically 1 to 5 cm in size, but locally up to 12 cm, and display local zoning and resorbed margins (Fig. GS-26-6e). Pyroxene glomerocrysts in subunit 10d are up to 4 cm in diameter. Thicker flows are locally gabbroic textured (subunit 10c) and include rare irregular pegmatitic domains up to 50 cm in diameter (Fig. GS-26-6f). A pegmatite pod from one thick flow was collected for U-Pb zircon dating in order to establish the age of the Gray Point sequence.

Intermediate to Felsic Volcaniclastic Rocks (Unit 11)

The base of the Drumming Point sequence is marked by an 80 to 120 m thick unit composed of bedded, intermediate to felsic volcaniclastic rocks. This unit is dominated by intermediate volcaniclastic rocks rich in plagioclase phenoclasts, but also contains chert, siliceous sedimentary rocks and heterolithic intermediate to felsic volcanic conglomerate, as well as rare rhyolite breccia and pillowed plagioclase-phyric basaltic andesite flows.

Intermediate volcaniclastic rocks (subunit 11a) are composed of 10 00 cm thick beds that display normal size-grading, basal scours, ma AB bed zoning, and rare flame structures and rip-ups (Fig. 26-7a). They are typically composed of texturally immature sand grit that is geochemically similar to the stratigraphically overlying nming Point andesite and basaltic andesite (unit 12). They most y represent locally derived, unconsolidated volcanic detritus that ıped and was deposited subaqueously by turbidity currents.

Intercalated with the intermediate volcaniclastic turbidites are beds olcanic conglomerate and breccia (subunit 11b). They form beds that e from 1 m to more than 4 m in thickness, vary widely in composi- and texture, contain a heterolithic clast population including intrusive rocks, are matrix-supported, and locally display both reverse and normal size-grading. Pebbles and cobbles in these beds are generally rounded to well rounded (Fig. GS-26-7b), in contrast to the detrital matrix, which normally is texturally immature. Many of the pebbles and cobbles are rhyolite and dacite, rock types not present in either the underlying (Gray Point) or overlying (Drumming Point) volcanic sequences. Rounding of pebbles and cobbles, heterolithic clast populations and presence of intrusive clasts suggest unroofing of the source terrane, possible subaerial transport of clasts, and mixing prior to their incorporation in the volcanic conglomerate beds. The texturally immature character of the detrital matrix indicates that it likely had a more local derivation and only limited, if any, subaerial transport.



ıre GS-26-7: Photographs of rocks in the Drumming Point sequence: a) normal size-grading of rhyolite pebbles in bed of intermedi- volcanic sandstone (unit 11); b) rounded rhyolite cobble supported in a matrix of texturally immature andesitic sandstone (unit 11); nfoliated dyke, texturally similar to Drumming Point andesite flows, cutting foliated Gray Point basalt (unit 9); d) pillowed Drumming t andesite (unit 12); e) F_2-folded, delicately laminated siltstone and mudstone (unit 13c); f) coarse greywacke with angular lithic frag- ts (unit 13b).

Chert beds occur sporadically throughout unit 11, suggesting periodic sediment starvation of the depositional basin. The presence of soft-sediment slumping of chert in one locality suggests basin instability during lithification.

The presence of this basal volcaniclastic unit at the contact between two geochemically distinct volcanic sequences, the MORB-like Gray Point and the calc-alkaline Drumming Point, suggests that these units may be separated by an unconformity or disconformity. The presence of subaerially transported, rounded pebbles and cobbles in conglomerate beds of unit 11, and unfoliated basaltic dykes, petrographically similar to Drumming Point volcanic rocks and cutting foliated unit 9 basalt on Gray Point (Fig. GS-26-7c), is consistent with this interpretation.

Pillowed Basaltic Andesite and Andesite (Unit 12)

Aphyric (subunit 12a) and, less commonly, plagioclase-phyric (subunit 12b; <10% phenocrysts, 0.5–3 mm in size) basaltic andesite and andesite flows form a 350 to 1100 m thick sequence that directly overlies volcaniclastic rocks of unit 11. The flows are buff to buff-brown on weathered surfaces and medium to pale grey-green on fresh surfaces. They are least deformed (subunit 12a) on islands directly north of Black Island. Most of the flows are pillowed (Fig. GS-26-7d) and include only rare massive types (subunit 12c). Flow thicknesses are generally not well constrained due to incomplete exposure of flows on small islands. Pillow selvages are typically less than 1 cm and interpillow hyaloclastite is typically between 1 and 4 cm thick. The flows contain rare amygdales, radial pipe vesicles and synvolcanic dykes. Interflow sedimentary deposits, which are common between Gray Point massive flows (unit 10), are absent in unit 12 flows. Pillowed flows on islands east of Gray Point are commonly strongly foliated, and many of the flows have been tectonized (subunit 12d), with lensoid lithons in which strongly flattened pillow relicts are preserved.

Feldspathic Wacke and Related Volcaniclastic Rocks (Unit 13)

The Drumming Point sequence is topped by up to 1000 m of feldspathic wacke and associated volcaniclastic rocks (unit 13). The thickness of this unit is not well constrained, as its eastern contact is unexposed and may be a D_1 fault.

The sequence is dominated by feldspathic wacke (subunit 13a) and a massive, medium-grained, feldspathic volcaniclastic equivalent (subunit 13b). Both subunits 13a and 13b are characterized by buff to pale buff weathering and a medium grey-green fresh surface. Deformed equivalents are typically pale grey-green on both weathered and fresh surfaces. Orangy pink feldspars, likely saussuritized and hematized, usually form between 50 and 75% of both these subunits, with the remainder consisting of amphibole and 5 to 20% quartz. The feldspathic wacke (subunit 13a) is typically well laminated but, because of its very limited variation in grain size, rarely forms well defined beds. It locally displays normal size-grading, scours and flame structures, and contains rare fine-grained lithic pebbles. Locally, it is intercalated with fine-grained siltstone, chert and mafic mudstone (subunit 13c; Fig. GS-26-7e). Crystal-rich feldspathic volcaniclastic rocks (subunit 13b) are typically massive, without discernable layering or bedding. Subunit 13b does, however, display local bedding and rare normal size-grading, attesting to its clastic character (Fig. GS-26-7f). A decrease in grain size from northern exposures of coarser, feldspathic volcaniclastic rocks (subunit 13b) to southern exposures of finer-grained, bedded, felspathic wacke (subunit 13a) suggests that rocks to the north may be more proximal deposits.

Mafic Mudstone (Subunit 13d)

Mafic mudstone (subunit 13d) outcrops on two small islands at the top of the exposed Black Island assemblage. The mudstone is dark grey-green on both weathered and fresh surfaces and characterized by a well developed parting. The mudstone may indicate upward fining toward the top of unit 13, but this is very speculative because of the wide spacing of exposures.

Rhyolite Breccia (Unit 14)

Rhyolite tuff–lapilli tuff (subunit 14a) and heterolithic felsic breccia (subunit 14b) are exposed on a single outcrop on the south shore of Black Island. This unit is significant, as Ermanovics and Wanless (1983) reported a U-Pb zircon age of 2732 ± 10 Ma for this outcrop, and suggested that the Black Island assemblage is equivalent to the Bidou Lake subgroup of similar age, exposed 65 km to the east at Bissett. However, fieldwork this summer shows that the rhyolite (unit 14) is in fault contact and faces opposite to the remainder of the Black Island assemblage (units 9–13). Thus, the age of units 9 to 13 is not known, and the suggested age correlation with the Bidou Lake subgroup at Bissett should be regarded as tentative.

Rhyolite tuff–lapilli tuff (subunit 14a), which forms the north side of the single outcrop of unit 14, is light grey to light buff-grey on weathered surfaces and pale grey on fresh surfaces. This greater than 5 m thick unit of aphyric, unbedded, massive rhyolite breccia mainly comprises a mixture of pale to medium grey-weathering, angular, fine-grained (?glassy) clasts and red-brown–weathering, subrounded, finely vesicular aphyric clasts. The presence in this mainly monolithic rhyolite breccia of possible shards and flattened pumiceous clasts indicates that it may have been a pyroclastic or secondary (resedimented) pyroclastic deposit.

Rhyolite cobble conglomerate (subunit 14b), which forms the south side of the outcrop, is well bedded and composed of a heterolithic mixture of subangular to rounded, matrix-supported felsic clasts. Beds vary in thickness from 30 cm to more than 100 cm, and one displays well defined normal size-grading of both matrix and larger clasts (Fig. GS-26-8a). A block of rhyolite tuff breccia in this conglomerate contains shards and welded pumice clasts (Fig. GS-26-8b), suggesting that at least part of the source volcanic terrane may have been subaerial. Rounded clasts and a heterolithic clast population are consistent with subaerial transport and mixing.

Gabbro, Quartz Diorite (Unit 15)

Quartz diorite to gabbro (unit 15) are exposed on a number of small islands to the south and east of Deer Island. Krogh et al. (1974) reported a 2715 ± 10 Ma U-Pb zircon age on a quartz diorite from this unit. The diorite and quartz diorite comprise fine- to medium-grained, multiphase intrusions (dyke complexes) that are overprinted by a weak ($?S_1$) foliation. Chilled margins and rare quartz amygdales in some dykes suggest intrusion at shallow levels and possible subvolcanic character. Two samples of unit 15 gabbro have chemical signatures identical to those of the Drumming Point basaltic andesite and andesite (unit 12).

GEOCHEMISTRY OF VOLCANIC ROCKS, BLACK ISLAND ASSEMBLAGE

Sampling of the Black Island assemblage volcanic rocks was undertaken in 1999 and 2000 as part of a broader program to provide high-precision geochemical analyses (ICP–MS) for volcanic rocks of the Rice Lake greenstone belt. The purpose is to characterize the trace-element and rare-earth element geochemistry of volcanic rocks in order to identify the tectonomagmatic environment of volcanism and to aid correlation between volcanic sequences throughout the belt (Bailes, 1998, 1999).

Samples were collected from all major units in the Black Island assemblage. They were taken from rocks that appeared mesoscopically least altered. Samples were trimmed to remove fractures, veins and weathered surfaces, and to avoid visible alteration or significant amygdale populations.

When this report was written, the only analyses available were

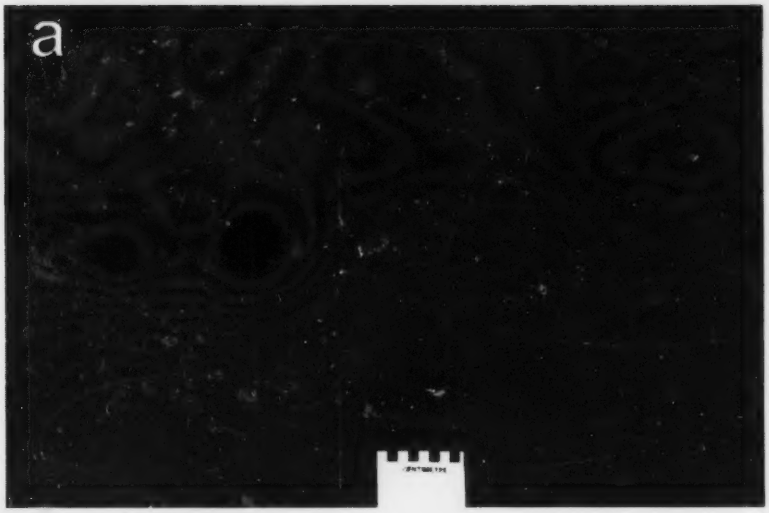


Figure GS-26-8: Photographs of rhyolite breccia (unit 14): a) normal size-grading at base of rhyolite pebble conglomerate (subunit 14a); b) rhyolite fragment containing flattened pumice and shards (subunit 14b).

those from the 1999 season, four from the Gray Point sequence and ten from the Drumming Point sequence. The stratigraphic positions of these sample sites are plotted for reference in Figure GS-26-5.

Major-element and immobile trace-element discrimination diagrams indicate that the Gray Point sequence (Fig. GS-26-9, a, c and e) is composed of basalt and basaltic andesite (46–54% SiO_2), whereas the Drumming Point sequence (Fig. GS-26-9, b, d and f) is slightly more felsic and composed of basaltic andesite and andesite (55–63% SiO_2). Both sequences are subalkaline (Fig. GS-26-9).

Gray Point basalt and basaltic andesite display many of the geochemical characteristics of ocean-floor basalt. They plot in the field of modern ocean-floor basalt on a diagram of Cr versus Ti (Fig. GS-26-10a) and in the N-MORB field using the trace elements Th, Zr and Nb (Fig. GS-26-10c). However, on diagrams using Th values or Th/Nb ratios (e.g., Fig. GS-26-11a, b), they plot within the arc field, slightly outside the ocean-floor/N-MORB field. Multi-element N-MORB–normalized plots (Fig. GS-26-12a) show that the basalt has elevated Th and slightly negative Nb values, but typical N-MORB geochemistry in all other respects. The elevated Th could be a result of crustal contamination, or the result of a slight subduction signature. Absence of negative values of Zr, Hf and Ti, which characterize subduction-arc–produced basalt in modern tectonic settings, suggests that the elevated Th may reflect contamination rather than a subduction component. An alternative interpretation, proposed for the geochemically similar Big Island basalts at Wallace Lake, is that this type of basalt was deposited in an oceanic plateau environment (Sasseville and Tomlinson, 2000). On primitive-mantle–normalized multi-element plots, the Gray Point basalt (Fig. GS-26-13a) displays slightly negative Nb and Th values and is geochemically similar to the Unnamed, Tinney, and Gunnar basalt units of the eastern Rice Lake greenstone belt. These outcrop 80 km to the east, where they form a basaltic 'platform' at the base of the Bidou Lake subgroup.

The Drumming Point basaltic andesite and andesite straddle the ocean-floor and island-arc fields of modern volcanic rocks on the Cr versus Ti diagram (Fig. GS-26-10b), but clearly fall in the arc field on the plot of Th, Zr and Nb (Fig. GS-26-10d) and on all discriminant plots shown in Figure GS-26-11. On multi-element N-MORB–normalized (Fig. GS-26-12b) and primitive-mantle–normalized (Fig. GS-26-13c, -13d) plots, the Drumming Point sequence displays negative Nb relative to Th, La and Ce, a feature that is typically attributed to subduction processes in modern volcanic regimes (Pearce and Peate, 1995). However, the positive normalized values of Zr and Hf on these plots are not typical of magmas in modern arc environments.

Elevated light rare-earth elements (LREE) and depleted heavy rare-earth elements (HREE) on the N-MORB– and primitive-mantle–normalized multi-element plots (Fig. GS-26-12b, GS-26-13c, -13d, -13e)

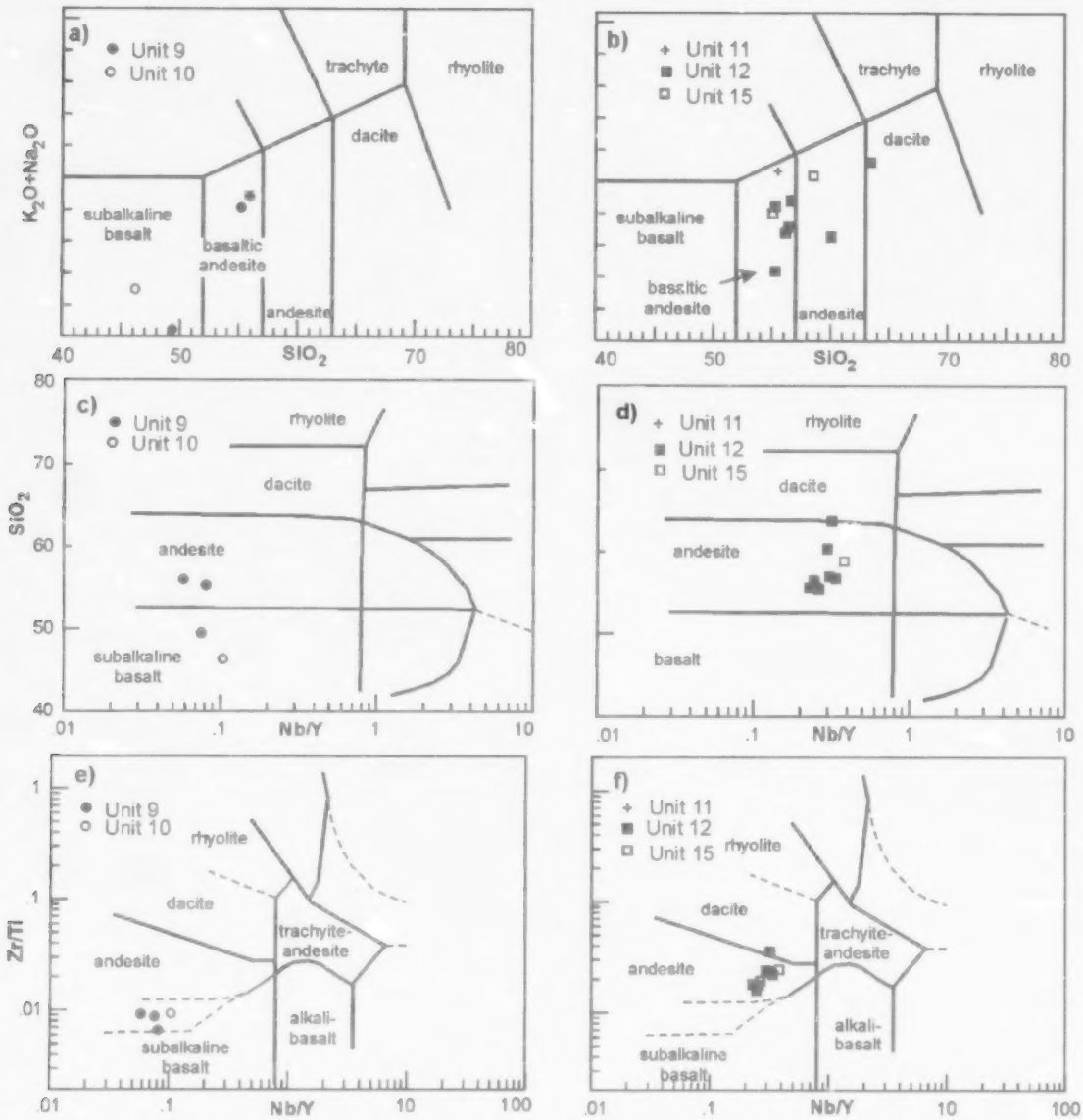


Figure GS-26-9: Black Island assemblage volcanic rocks plotted on geochemical discrimination diagrams for rock type and major chemical affinities: SiO_2 vs. K_2O+Na_2O (a and b), Nb/Y vs. SiO_2 (c and d), and Nb/Y vs. Zr/Ti (e and f).

suggest a calc-alkalic affinity for the Drumming Point sequence. Drumming Point volcanic rocks also fall in the calc-alkalic field on the Jensen cation plot (Fig. GS-26-14b) and display Hf/Th ratios between 1.3 and 2.4. The latter is significant, as Wood (1980) has shown that modern tholeiitic and calc-alkalic volcanic suites have Hf/Th ratios of greater than 3 and less than 3, respectively. The Drumming Point sequence is geochemically similar to basaltic andesite and andesite that comprise part of the upper Bidou Lake subgroup at Bissett and Wanipigow Lake (Fig. GS-26-13f).

The dramatic geochemical differences between the Gray Point and Drumming Point sequences, N-MORB (?contaminated) versus calc-alkaline, indicate that the boundary between these two volcanic sequences, which appears to be marked by an unconformity, also marks a significant change in tectonic setting of volcanism. The geochemical similarity of Gray Point and Drumming Point volcanic rocks to the lower and upper portions of the Bidou Lake subgroup, respectively, suggests that this boundary could extend to the east into the Bissett and Beresford Lake areas.

ECONOMIC GEOLOGY

The Rice Lake greenstone belt has a long history as a lode gold producer. Shear zones and related quartz-carbonate vein stockworks, belonging to D_1, on the eastern shore of Black Island and on smaller islands to the east are considered to have potential to host gold, but assayed samples from two sites showed less than 8 ppb Au. However, very little exploration activity (trenches or sampling) was obvious on these zones, indicating that their potential has not been assessed.

The Rice Lake greenstone belt hosts no known economic base-metal volcanogenic massive sulphide (VMS) deposits. Rhyolite in the Rice Lake greenstone belt is typically F2 type, according to the classification system of Lesher et al. (1986), and therefore does not have a high potential for economic VMS deposits (Lesher et al., 1986). The absence

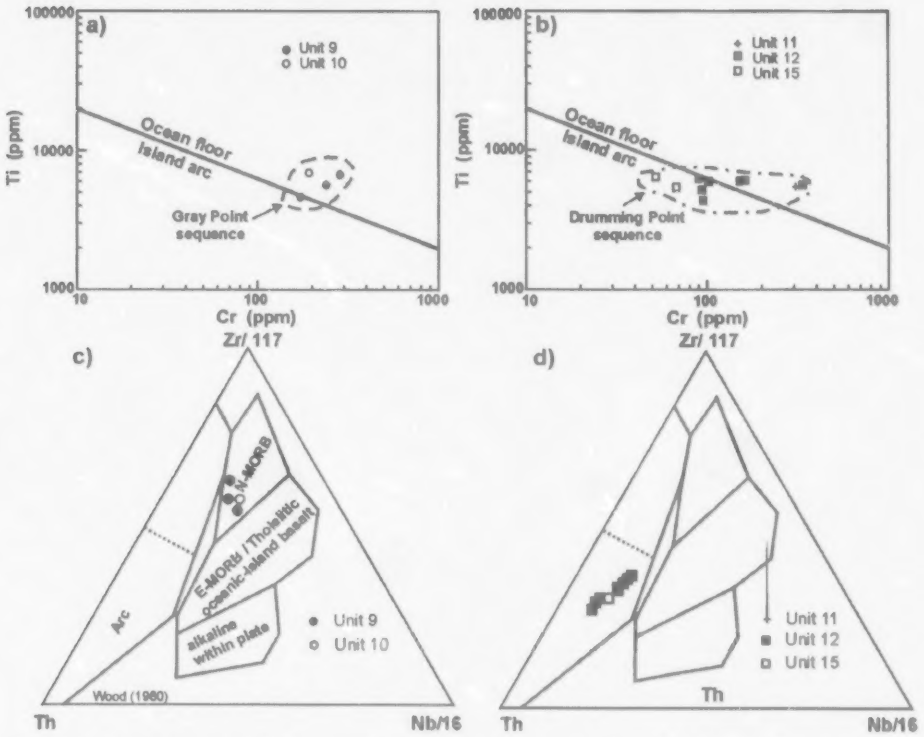


Figure GS-26-10: Black Island assemblage basalt, basaltic andesite and andesite plotted on tectonic discrimination diagrams: Cr vs. Ti (a and b); Th vs. Zr/117 vs. Nb/16 (c and d).

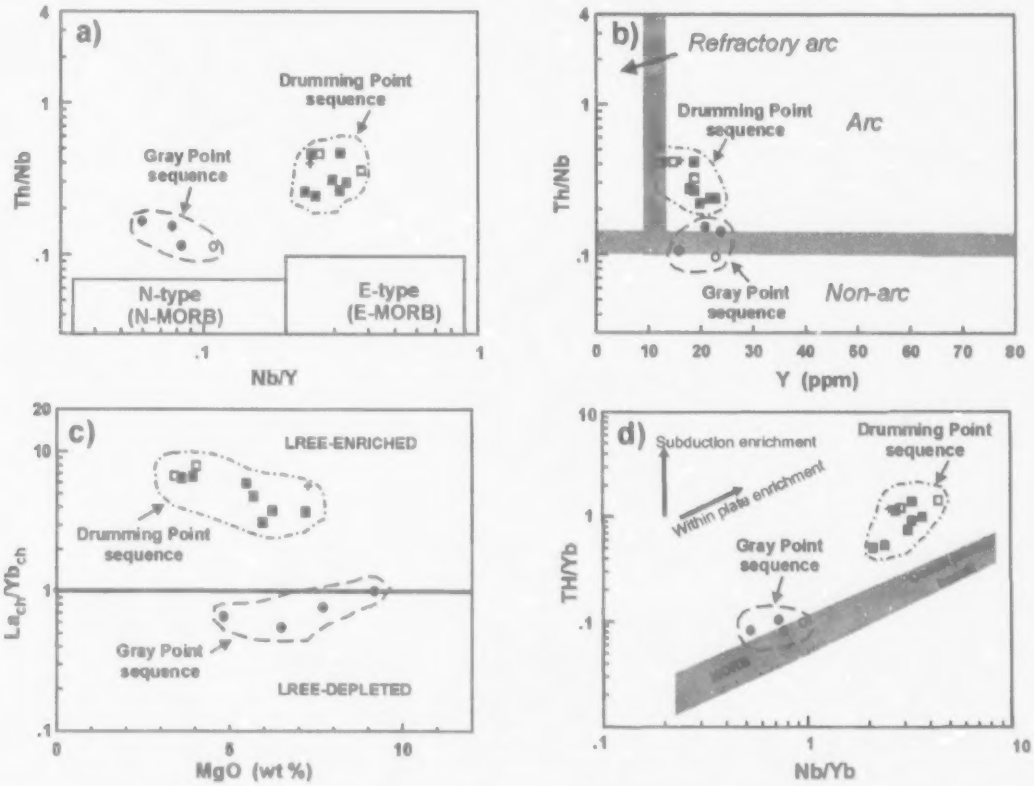


Figure GS-26-11: Black Island assemblage volcanic rocks plotted on various geochemical discrimination diagrams: a) Nb/Y vs. Th/Nb, b) Y vs. Th/Nb, c) MgO vs La_{ch}/Yb_{ch} and d) Nb/Yb vs. Th/Yb.

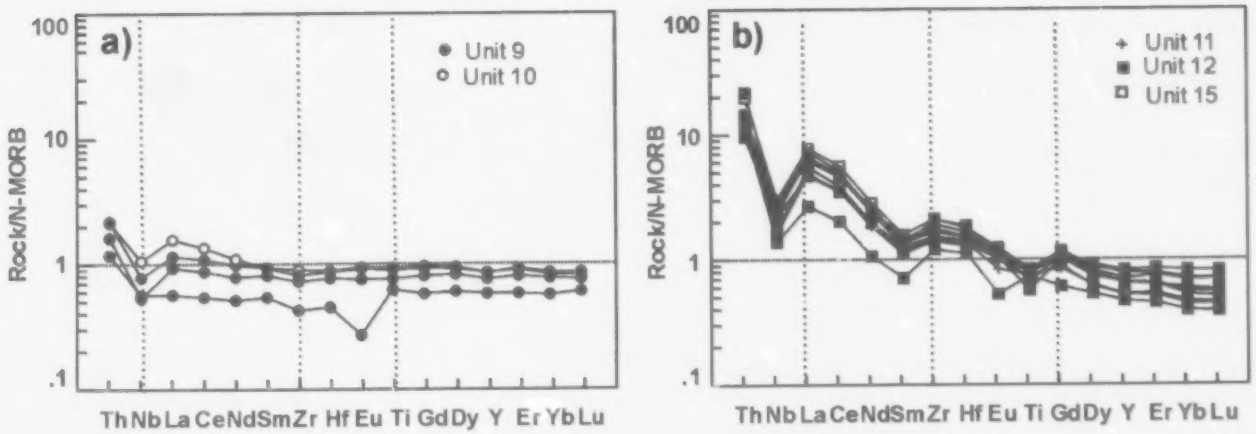


Figure GS-26-12: N-MORB–normalized multi-element plots of Black Island assemblage mafic volcanic rocks. Normalizing values and order adopted from Sun and McDonough (1989).

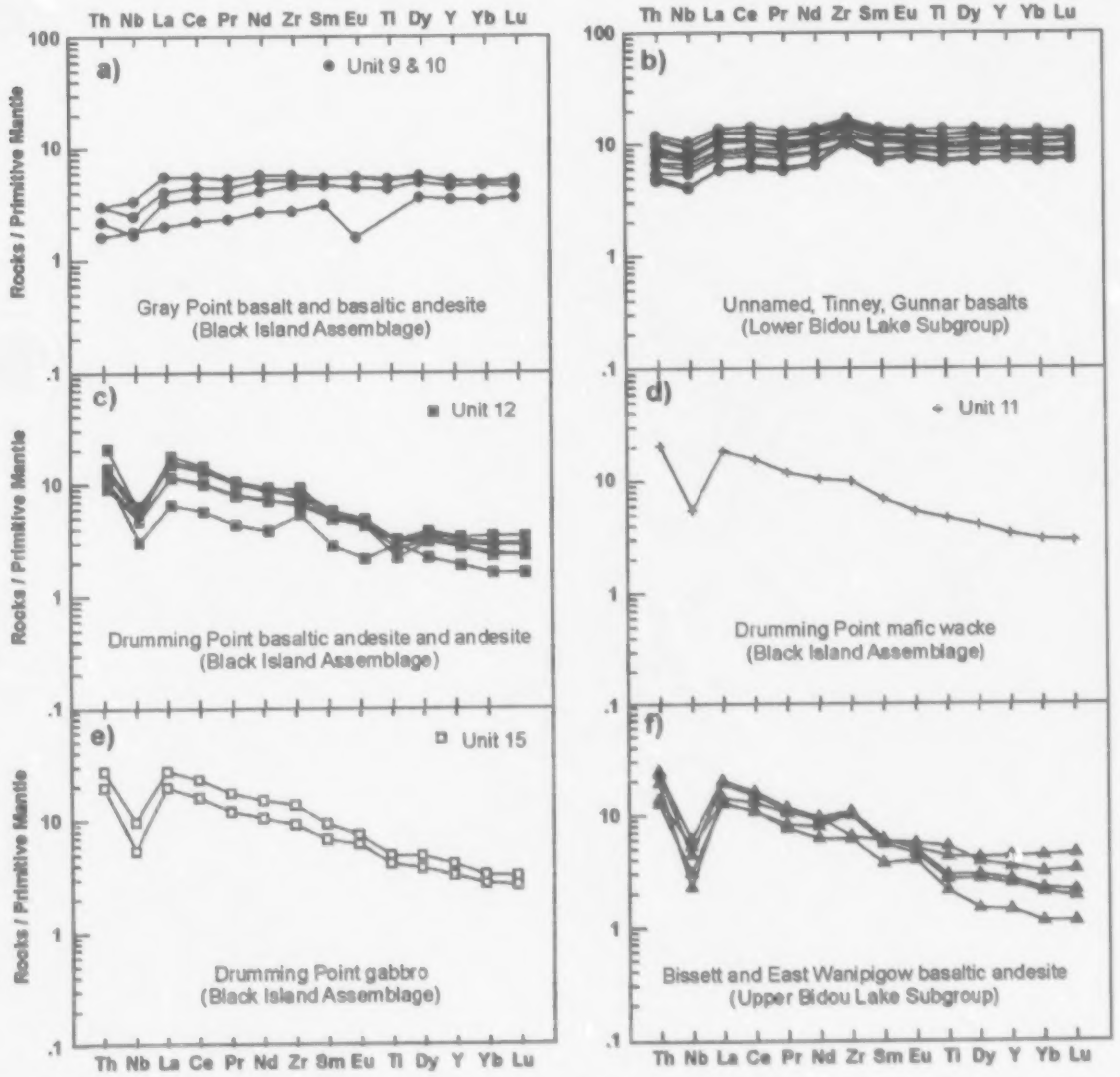


Figure GS-26-13: Primitive-mantle–normalized multi-element plots of Black Island assemblage and Bidou Lake subgroup mafic volcanic rocks. Normalizing values and order adopted from Sun and McDonough (1989).

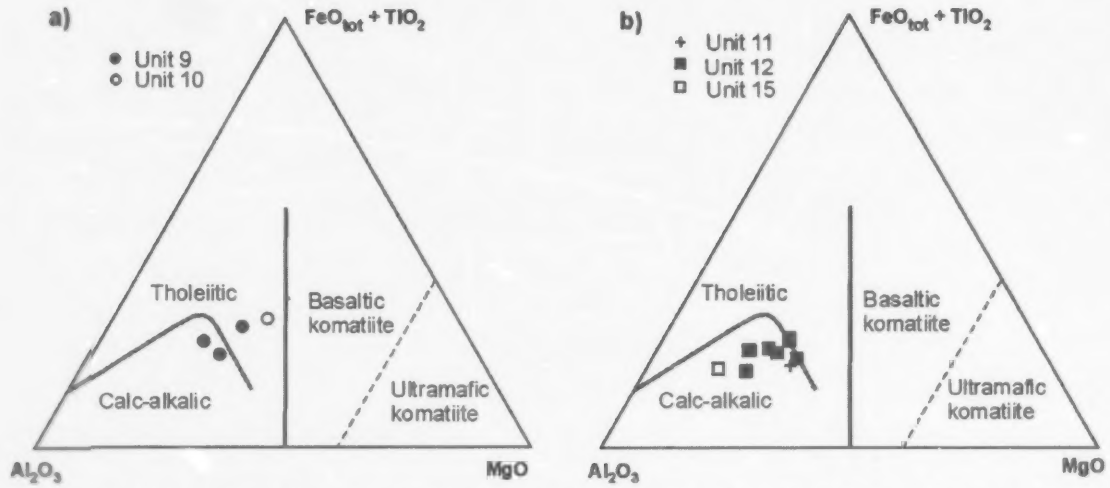


Figure GS-26-14: Black Island assemblage volcanic rocks plotted on Jensen cation plot (Al_2O_3 vs. $FeO_{tot}+TiO_2$ vs. MgO): a) Gray Point basalt and basaltic andesite; b) Drumming Point basaltic andesite and andesite.

of rhyolite in the Black Island assemblage (only two small exposures), combined with the F2 geochemical signature of rhyolite in the Rice Lake belt, suggests that copper-zinc VMS deposits are unlikely to be present in the Black Island assemblage.

The presence of ca 3.0 Ga crust (units 1 and 2) east of Lake Winnipeg and the unconformably overlying Lewis–Storey rift-platform sedimentary rocks ((units 3–7) provides a previously unrecognized and unexplored succession. The ca. 3.0 Ga ([?] and older) crust is one that could have potential for diamond exploration, since old Archean crust can be bonded to stable refractory mantle lithosphere that includes the diamond stability field (Kerrich and Wyman, 1996). The economic potential of the Lewis–Storey rift-platform sequence is not known. However, rifted continental margins elsewhere are commonly areas with elevated levels of hydrothermal activity and ultramafic magmatism. Younger rift successions are known to host magmatic nickel and platinum-group elements (PGEs).

CONCLUSIONS

The succession of rocks exposed on Black Island and the adjacent east shore of Lake Winnipeg provides a unique opportunity to examine the tectonostratigraphic and structural relationships between the Uchi Subprovince (Black Island assemblage) and the older North Caribou Terrane. A working hypothesis for the tectonostratigraphic and structural evolution of the area involves deposition of the Lewis–Storey assemblage as a rift sequence on the western margin of ca. 3 Ga basement (East Shore plutonic complex), with subsequent spreading recorded by mafic sill emplacement. The deposition of the Black Island assemblage began in an ocean-floor–arc-rift environment (Gray Point sequence, N-MORB geochemistry) and abruptly switched ([?] at a disconformity) to an oceanic-arc–like environment (Drumming Point sequence, calc-alkalic). Juxtaposition of the Black Island assemblage and the North Caribou Terrane (at ca. 2.71–2.70 Ga) resulted in development of D_1 structures in a probable strike-slip (dextral) regime. The Hole River sedimentary rocks were deposited in strike-slip basins during continued D_1 oblique convergence. Waning deformation became focused in shear zones during D_2 and D_3 transcurrent shear events.

The new mapping and interpretation of rocks undertaken as part of this project provide a more complete geological context in which to undertake exploration at the west end of the Rice Lake greenstone belt. The unique rock formations and complex overprinting deformation provide the possibility of mineralization in stratigraphic and structural environments that have yet to be satisfactorily explored and investigated.

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GS-27 NEW PGE-RELATED RESULTS OF STUDIES ON THE CHROME AND PAGE PROPERTIES, BIRD RIVER SILL (NTS 52L/5)

by P. Theyer

Theyer, P. 2000: New PGE-related results of studies on the Chrome and Page properties, Bird River Sill (NTS 52L/5); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p.175-178.

SUMMARY

The Bird River Sill has been explored for chromite, Ni-Cu and platinum group elements (PGEs). A recent unprecedented increase in both demand for and price of PGEs, plus the demonstrated potential of the Bird River Sill to contain economic PGE deposits, have led to renewed interest in exploring the sill for these minerals. A sulphide- and PGE-bearing rock layer hosted by a highly deformed peridotite unit was defined, mapped and sampled on the Chrome property. Two PGE-bearing layers associated with chromitite had previously been defined on the property. Field investigations show that the mineralized layers are stratabound but disrupted by block faulting.

A sulphide-bearing peridotite layer, associated with chromitite in a stratigraphic position analogous to those on the Chrome property, was defined on the Page property, located approximately 2.5 km northeast of the Chrome property.

INTRODUCTION

The Bird River Sill is an approximately 20 km long, east-striking, mafic-ultramafic complex of Archean age (2745 ± Ma; Timmins et al., 1985) intrusive into the supracrustal Lamprey Falls and Bernic Lake formations of the Rice Lake Group. Subsequent granitic intrusions associated with deformation and complex faulting segmented the originally contiguous sill into offsets that, in places measure more than 1.5 km and have obliterated and/or duplicated parts of the sill (Fig. GS-27-1).

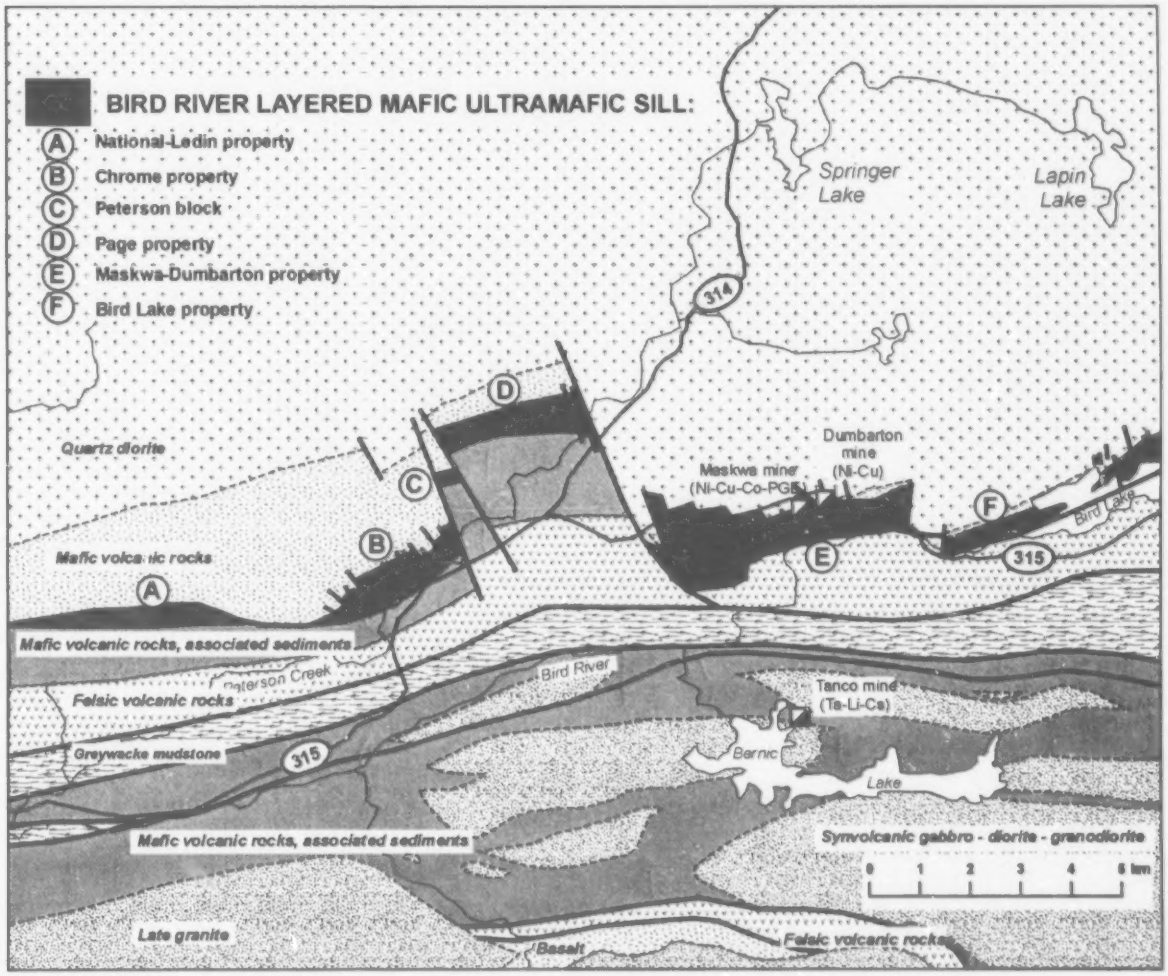

Figure GS-27-1: Geology of the southern part of the Bird River greenstone belt, southeastern Manitoba, showing locations and names of properties on the Bird River Sill (modified from *'Geological and proposed work report for the Page property', unpublished internal report, Canmine Resources Corporation, 2000).*

Geological investigations in the 1940s and early 1950s concentrated on the search for chromite. Exploration focused on Ni-Cu mineralization in the late 1950s, but interest reverted to chromite in the early 1980s. In the late 1980s, government-sponsored studies concentrated on the sill's chromite reserves and the feasability to concentrate and upgrade these for use as mill feed in the production of stainless steel.

Investigations of the sill's potential to contain PGEs started in 1982 when the author cut and analyzed a nearly continuous channel sample across the sill on the Chrome property (Theyer, 1982). Inclusions of PGEs in chromite from the Bird River Sill were described by Talkington et al. (1983). Theyer (1985, 1991) described two sulphide layers, each several metres thick, intersected in the ultramafic portion of his channel samples. Scoates et al. (1989) published maps and descriptions of two additional PGE-bearing layers, one associated with chromitite of the Lower Group and the other with chromitite of the Disrupted Group.

Peck and Theyer (1998) re-analyzed Theyer's (1985) channel sample material and re-affirmed that two stratigraphic intervals (34–52 m and 95–115 m, western cut; Theyer, 1985, Fig. GS-27-10), in which sulphide minerals and PGEs were concentrated, occur stratigraphically below the layers of the chromitiferous zone (Fig. GS-27-2). One of these layers coincides with the PGE-bearing rock layer described by Scoates et al. (1989).

CHROME PROPERTY

Sulphide- and PGE-Bearing Disrupted (Foliated) Peridotite Layer

The sulphide-bearing disrupted (foliated) peridotite layer is in the basal subzone of the layered zone (suite) of the ultramafic series (Scoates, 1983) of the Bird River Sill (Fig. GS-27-2 and -3). Host rocks to this subzone are disrupted, foliated and incompetent peridotite,

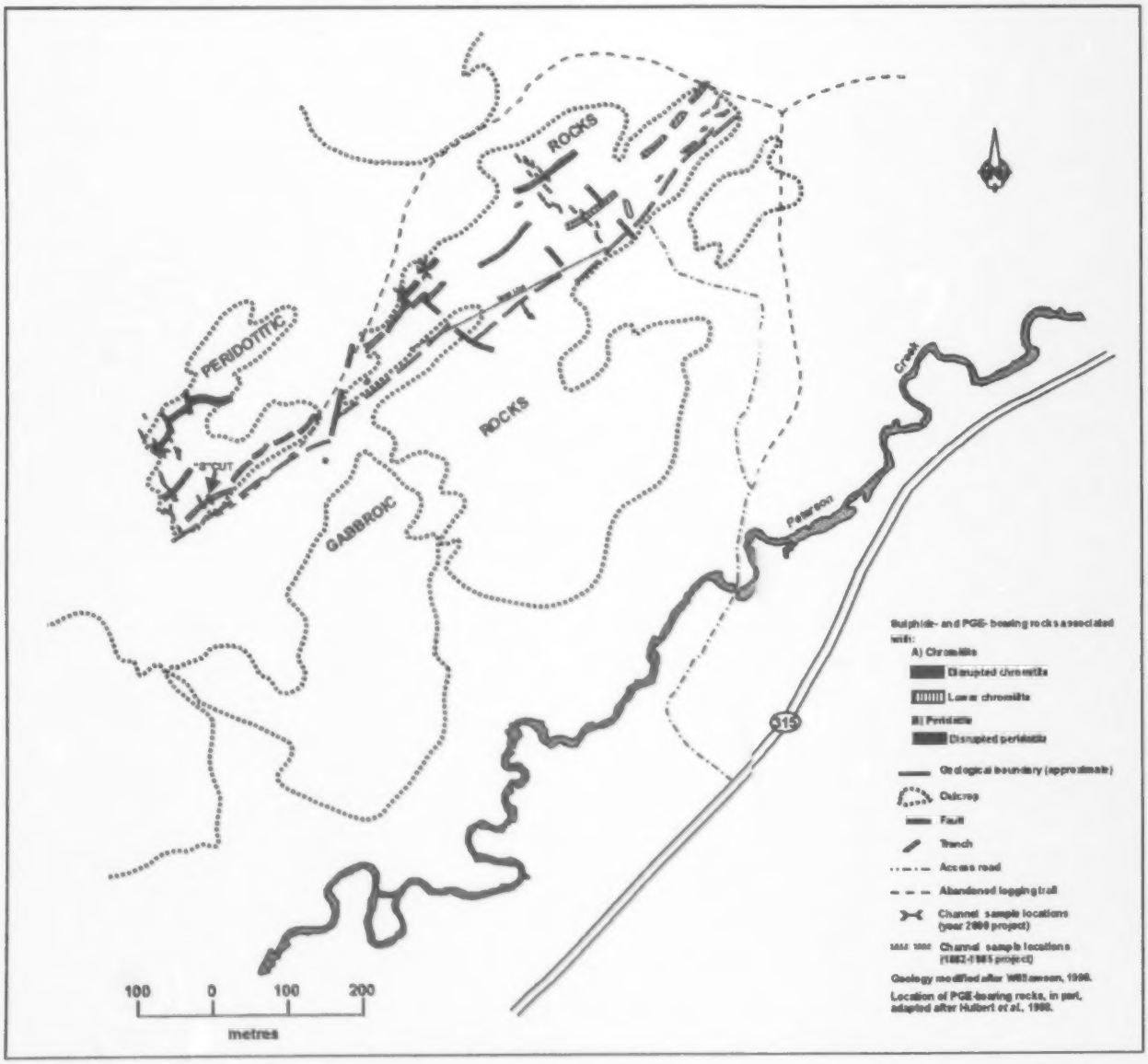

Figure GS-27-2: Geology of the Chrome property, showing location and extent of chromitite layers and of sulphide- and PGE-bearing layers. Geology and location of some sulphide bearing layers from Williamson (1990) and Hulbert et al. (1988).

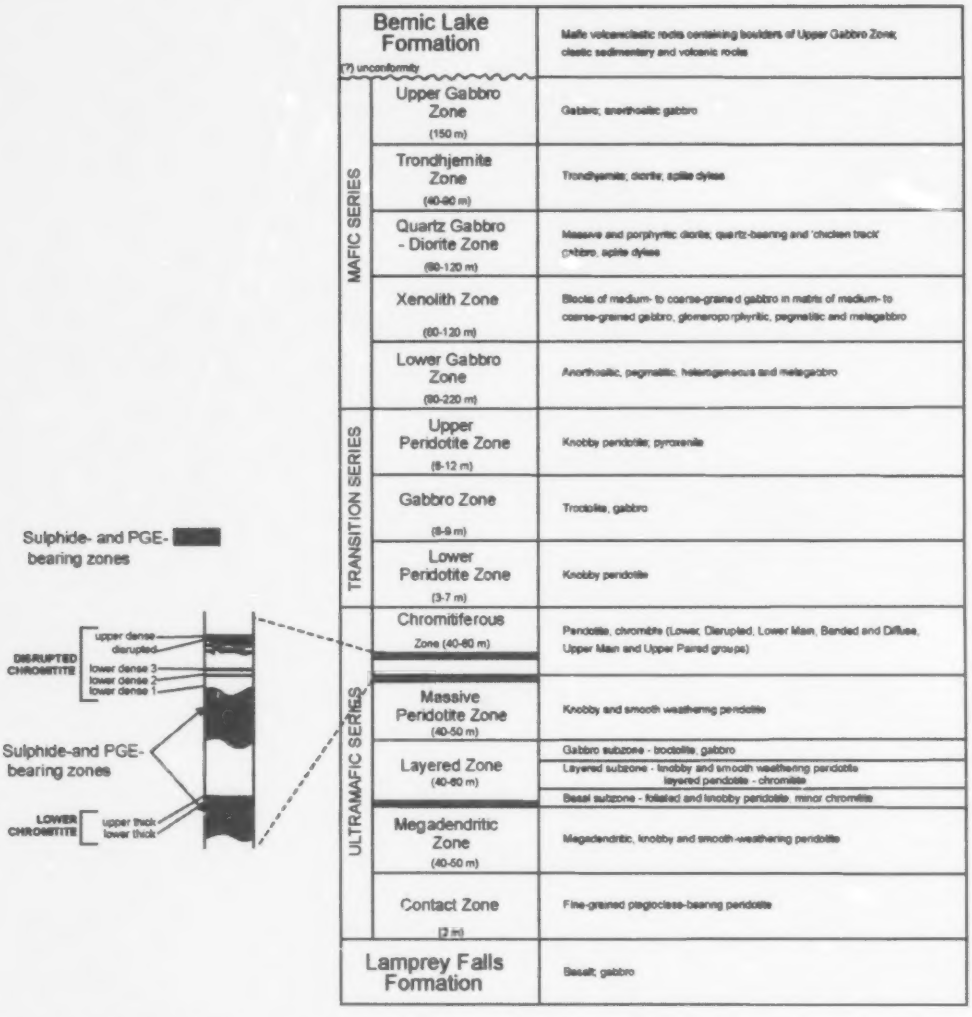
Bernic Lake Formation
Mafic volcaniclastic rocks containing boulders of Upper Gabbro Zone; clastic sedimentary and volcanic rocks
(?) unconformity
MAFIC SERIES
Upper Gabbro Zone (150 m)
Gabbro; anorthositic gabbro
Trondhjemite Zone (40-90 m)
Trondhjemite; diorite; aplite dykes
Quartz Gabbro - Diorite Zone (60-120 m)
Massive and porphyritic diorite; quartz-bearing and 'chicken track' gabbro; aplite dykes
Xenolith Zone (60-120 m)
Blocks of medium- to coarse-grained gabbro in matrix of medium- to coarse-grained gabbro, glomeroporphyritic, pegmatitic and metagabbro
Lower Gabbro Zone (60-220 m)
Anorthositic, pegmatitic, heterogeneous and metagabbro
TRANSITION SERIES
Upper Peridotite Zone (8-12 m)
Knobby peridotite; pyroxenite
Gabbro Zone (8-9 m)
Troctolite, gabbro
Lower Peridotite Zone (3-7 m)
Knobby peridotite
ULTRAMAFIC SERIES
Chromitiferous Zone (40-60 m)
Peridotite, chromitite (Lower, Disrupted, Lower Main, Banded and Diffuse, Upper Main and Upper Paired groups)
Massive Peridotite Zone (40-50 m)
Knobby and smooth weathering peridotite
Layered Zone (40-60 m)
Gabbro subzone - troctolite, gabbro
Layered subzone - knobby and smooth weathering peridotite layered peridotite - chromitite
Basal subzone - foliated and knobby peridotite, minor chromitite
Megadendritic Zone (40-50 m)
Megadendritic, knobby and smooth-weathering peridotite
Contact Zone (2 m)
Fine-grained plagioclase-bearing peridotite
Lamprey Falls Formation
Basalt; gabbro
Sulphide- and PGE-bearing zones
DISRUPTED CHROMITITE
upper dense disrupted
lower dense 3
lower dense 2
lower dense 1
Sulphide-and PGE-bearing zones
LOWER CHROMITITE
upper thick
lower thick

*Figure GS-27-3: Stratigraphy of the Bird River Sill, its immediate host rocks and sulphide- and PGE-bearing layers in the area of the Chrome property (*modified from *Scoates et al., 1989).*

characterized by a rust-stained, rough weathered surface. Exposed widths range from 5 to 16 m. The contact with the stratigraphically underlying, mechanically competent rock of the 'megadendritic zone' in places consists of a prominent and abrupt ledge, up to 2 m high, likely caused by relative differences in the resistance to weathering and erosion of the two lithological units. Sulphide mineralization consists mainly of subrounded, crystal-supported pyrrhotite droplets of distinctly magmatic appearance. Subordinate amounts of chalcopyrite and pyrite occur as overgrowths on pyrrhotite. Both the concentration and the distribution of sulphides vary considerably (e.g. in places, there were 10–15 cm thick layers with up to 3% pyrrhotite but also subrounded decimetre-sized accumulations of disseminated sulphides). Fractures that cut and offset the mineralized layer are, in places, characterized by centimetre-thick fault gouge, in places stained with blue and green blotches interpreted as products of the supergene alteration of chalcopyrite. The basal subzone is underlain by the megadendritic peridotites of the megadendritic zone (Williamson 1990). The peridotites of the megadendritic zone are characterized by decimetre-long, arcuate, dendritic pyroxene 'sprays' that impart an unmistakably ornamented appearance to the weathered surface. The disrupted peridotite layer is stratigraphically overlain by finely banded, layered peridotite devoid of sulphide minerals (Fig. GS-27-3).

Mineralization

Two channel samples were cut with a rock saw across parts of this layer and rock samples were sent for chemical analysis. Results of these analyses were not available at the time this report was written.

Cut 1 is a 5.5 m long channel sample sawn across the disrupted peridotite in the northwestern part of the Chrome exposure. Cut 2 is an approximately 6.5 m long channel sample (Fig. GS-27-2). Sulphide mineralization in both samples consists of up to 5% pyrrhotite in coarse-grained peridotite.

Sulphide- and PGE- Bearing Layers Associated with the Lower Group and Disrupted Group Chromitite

Two sulphide- and PGE-bearing layers, one associated with the Lower Group chromitite layer and the other with the Disrupted Group chromitite layer (Fig. GS-27-2; Scoates, 1983; (Scoates et al., 1989) were traced and sampled.

Each of the layers was sampled by means of metre-long channel samples located roughly in the centre of the Chrome property (cuts 3 and 4, Fig. GS-27-2). These samples were collected to complement the channel samples of the western and eastern Chrome property cut during the 1982–1985 project (Fig. GS-27-2).

Mineralization associated with the Lower Chromitite layer

Cut 3 (Fig. GS-27-2) is a 2.1 m long channel sample intersecting a fault-bound block of dense, fine-grained peridotite. The southern boundary of the mineralized zone, coinciding with the southern end of the cut, is 13.3 m north (stratigraphically below) of the lower thick member of the Lower Group chromitite. However, this distance may be of minor stratigraphic validity, since an intervening fault characterized by an approximately 5 cm thick mylonite layer is interpreted to indicate a relative displacement of these rock units.

Sulphide minerals consist of approximately 3% disseminated pyrrhotite, in places occurring as subrounded droplets interpreted to indicate their magmatic nature. Chalcopyrite was observed in trace quantities as fracture fillings.

Mineralization Associated with the Disrupted Chromitite Layer

Cut 4 is located approximately 1.4 m stratigraphically below the 'lower dense 2' (Scoates, 1983) member of the Disrupted Group chromitite layers (Fig. GS-27-2). The host rock is a coarse-grained dunite characterized by millimetre-sized amoeboid olivine crystals and centimetre-sized, subrounded, aphanitic, glassy, dark-coloured inclusions. Mineralization consists of approximately 4% pyrrhotite blebs ±chalcopyrite and abundant chromite in centimetre-sized pods and blebs. The weathered surface is characterized by abundant rusty and rare blue-green spots, attributed to supergene alteration of Fe and Cu sulphide minerals.

Analytical data for these samples had not been received by the time this report was written.

"S cut"

This approximately 5 m long channel sample, cut approximately 45 m east of the western cut on the western Chrome property in 1986 (Fig. GS-27-2), is outside the area mapped in detail by Williamson (1990). However, the stratigraphy and facies of the chromitite layers and the lithology of the host rocks suggest that this sample intersected the stratigraphically lower members of the Disrupted Group of chromitite layers.

Principal oxide minerals are magnetite, chromite and subordinate ilmenite. Sulphide minerals were identified as chalcopyrite, pyrite, pentlandite and millerite. High PGE concentrations (1600 ppb Pd, 610 ppb Ru and 390 ppb Pt) occur in an approximately 20 cm thick layer, that is particularly enriched in sulphide minerals and impoverished in chromite. This layer is located approximately 30 cm stratigraphically below an approximately 5 cm thick chromitite layer that was tentatively identified as the lower dense 1 member of the Disrupted Group of chromitite layers. Laurite (RuS_2), ocurring as idiomorphic inclusions in chromite, was the sole identifiable PGE mineral (T. Weiser, pers. comm., 1987).

PAGE PROPERTY

The Page property, located approximately 3 km northeast of the Chrome property (Fig. GS 29-1), is a faulted block of the Bird River Sill. Although its stratigraphy and lithology are comparable to those of the Chrome property, quality and quantity of outcrops and ease of access have tended to focus attention on the latter property. Mapping of the chromitite layers on the Page property (Young, 1992) showed a gross correspondence of the stratigraphy of the chromitite layers with those of the Chrome property.

Considering the close correlation between sulphide- and PGE-bearing layers and the Disrupted and Lower groups demonstrated for the Chrome property, it was considered likely that such a correlation could be established for the Page property. A preliminary field inspection of chromitite layers on the property showed that a sulphide-bearing rock layer occurs in the vicinity (10 m north) of chromitite layers tentatively identified as part of the Disrupted Layer Suite (Young 1992), which is equivalent to the Disrupted Layer Suite of the Chrome property (as defined by Scoates, 1983).

The approximately 1 m long channel sample intersected dense, fine-grained peridotite containing approximately 2% disseminated pyrrhotite. Analytical results were not available at the time this report was written.

RECOMMENDATIONS FOR PGE EXPLORATION IN THE BIRD RIVER SILL

In addition to previous recommendations made by Peck and Theyer (1998), it can be stated that:

1) the extent of an additional, formerly ill-defined stratabound sulphide- and PGE-bearing layer on the Chrome property should be carried out; and
2) the discovery of sulphide-bearing peridotite, associated with chromitite on the neighbouring Page property, is an indication that the Bird River Sill is an excellent target for PGE exploration.

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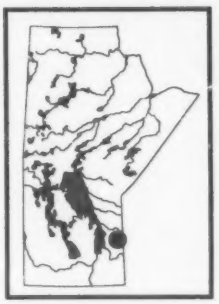
GS-28 PHYTOREMEDIATION AND PHYTOMINING IN MANITOBA: PRELIMINARY OBSERVATIONS FROM AN ORIENTATION SURVEY AT THE CENTRAL MANITOBA (AU) MINESITE (NTS 52L/13)



by S. Renault[1], E. Sailerova[2] and M.A.F. Fedikow

Renault, S., Sailerova, E. and Fedikow, M.A.F. 2000: Phytoremediation and phytomining in Manitoba: preliminary observations from an orientation survey at the Central Manitoba (Au) minesite (NTS 52L/13); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 179-188.

SUMMARY

A new project to establish the trace-element content of abandoned minesite tailings and determine the potential for phytoremediation of these sites and production of bio-ores, through the identification of hyperaccumulator plant species, has been initiated at the Central Manitoba (Au) minesite in southeastern Manitoba (Fig. GS-28-1). The first phase of this project has established the trace-element characteristics of tailings at the minesite through multi-element analysis of samples collected from three 1 m deep, hand-augered profiles. Results indicate that a partial extraction at pH 2–3 liberates a wide range of base and precious metals from the tailings at the minesite, and that these metals provide the focus for the future development of a bio-ore. Ten plant species were planted on three experimental sites on mine tailings. The survival rate and amount of extracted base and precious metals should indicate plant species suitable for phytoremediation and phyto-extraction.

INTRODUCTION

Mining and mineral-processing activities contribute significant quantities of heavy metals to the environment and affect surrounding land, air and water quality. These effects cannot be reversed by nature and require aggressive application of planned restoration programs.

Metal mines throughout Manitoba and Canada contain variable amounts of sulphide minerals, either in the ore or in the surrounding host rocks. Leaching of heavy metals from tailings results from oxidation of pyrite under moist conditions. This oxidation leads to a decrease in pH, which increases the solubility of most heavy metals. Metal mobilization

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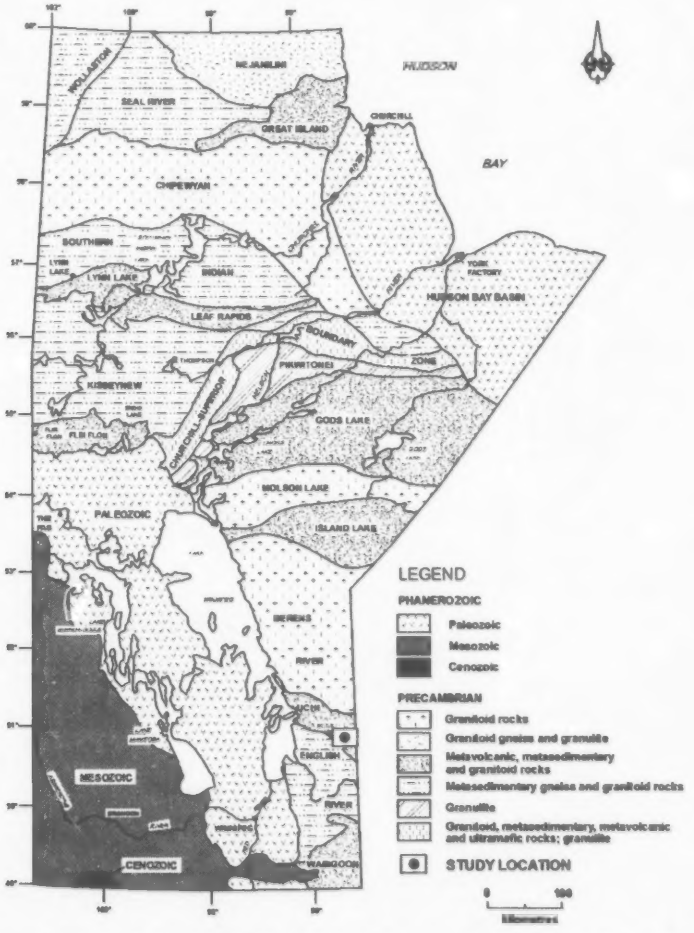

Figure GS-28-1: Location of the phytoremediation and phytomining study area, Central Manitoba gold mine, Bissett area, southeastern Manitoba.

from tailings can often be traced into watersheds downstream from active minesites.

Among the most promising and common approaches to preventing acid generation is to physically stabilize the waste by covering the acid-producing material with water. This forms an anaerobic environment, where sulphide minerals remain stable and dissolved metal concentrations are low. However, more than half of the existing tailings sites in Canada have been found to be unsuitable for water barriers (Ripley et al., 1996).

Pyrite oxidation can also be controlled by covering the surface tailings with soil and then usually revegetating the site. Establishment of a self-sustaining mat of vegetation is usually an important element in a rehabilitation program for waste-disposal areas. Vegetation stabilizes the soil, prevents new acid-generating material from being exposed, and decreases the amount of water available for deep percolation through transpirational water movement. Vegetation is established by controlling pH near the surface with addition of lime or limestone and by adding fertilizers where necessary. Plant species are usually selected from well adapted species generally growing in the region (Bradshaw, 1952). Nevertheless, revegetation alone does not stop acid drainage and does not remove heavy metals from the contaminated site.

Tailings are not the only cause of land contamination by heavy metals. Additional contamination is contributed by airborne pollutants, which are produced during mining and milling operations. Airborne pollutants are the major source of soil contamination in areas surrounding the mining operations. The distance from a source at which metals affect plants and soil has been documented. In Flin Flon, manganese was detectable for approximately 250 km from a base-metal smelter, and copper for up to 60 km. Affected soil cannot be used for agricultural purposes because of the high metal content, and other uses of these lands are also limited. The Subcommittee on Environmental Quality Criteria for Contaminated Sites of the Canadian Council of Ministers of the Environment has developed a set of interim criteria as part of the National Contaminated Sites Remediation Program. In some jurisdictions, these guidelines are being applied as limits for permitting purposes. Today's practices involve the expensive removal of contaminated soil (excavation; Giasson and Jaouich, 1998).

Phytoremediation of metals is being developed as a potential cost-effective remediation solution for thousands of sites contaminated by heavy metals in the United States and abroad (Salt et al., 1995, 1998; Cunningham et al., 1995; Comis, 1996). Plants, called hyperaccumulators, capable of accumulating 0.5 to 1% of their dry weight in metals have been identified. These include *Alyssum bertolonii* (1% or 10 000 mg/kg Ni) and *Brassica juncea* (Au). Anderson et al. (1999) provided a list of other hyperaccumulators.

Two basic strategies for phyto-extraction can be applied. The first of these is continuous phyto-extraction, which requires hyperaccumulators with high biomass production and the ability to grow in dense stands. Continuous phyto-extraction is also useful at sites where metals are already mobilized by low pH of the underlying soil, provided that the metals to be extracted are pH sensitive or contain high concentrations of metals in water-soluble form. The second approach is induced phyto-extraction (Blaylock et al., 1997). At the end of the plant-growth phase, the appropriate chelate, such as ethylene diamine tetra-acetic acid (EDTA), is applied to the area and plants are harvested within several days or a week. Chelate-assisted transport of metals to plant shoots appears to occur in the xylem via the transpiration stream. The metal appears to move to the shoots as a metal-chelate complex.

The technique of phytomining (Nicks and Chambers, 1998) involves growing a crop of a metal-hyperaccumulating plant species, harvesting the biomass, and burning it to produce a bio-ore. Up to 57 mg/kg Au (dry mass) can be accumulated by Indian mustard (*Brassica juncea*; Anderson et al., 1998) using induced hyperaccumulation by ammonium thiocyanate, which is biodegraded to ammonia, bicarbonate and sulphate.

Phytomining technology offers the possibility of exploiting ores or mineralized soils that are uneconomic by conventional mining methods. Bio-ores are virtually sulphur free, and their smelting requires less energy than sulphide ores. The metal content of a bio-ore is usually much greater than that of a conventional ore and therefore requires less storage space, despite lower density. Moreover, phytomining is an environmentally responsible approach to site remediation.

OBJECTIVES

The long-term goal of this study is to determine the optimal field conditions and define limiting factors for phytoremediation of sites contaminated with heavy metals. This field and laboratory study will guide the selection of suitable plant species for environmental conditions found in the boreal-forest region. The suitability of selected species for phytomining of base metals and gold will be tested in terms of the quality and costs of bio-ore production and economic effectiveness.

The initial experimental sites will be evaluated in terms of the degree and type of contamination, soil quality (including an estimation of soil pH and soil-layer thickness on revegetated areas), presence of other plant species, and soil drainage (Sailerova, in press). Furthermore, experimental sites must be evaluated for conditions that will enable the selected plant species to reach maximum biomass. Different plant species have different saturation irradiation levels and different requirements in terms of soil quality, mineral nutrition and water supply.

SIGNIFICANCE OF THE WORK

Soil contamination by heavy metals and metal leakage from sulphide tailings represent a significant and widely recognized ecological hazard. Phytoremediation offers the possibility of an ecologically acceptable and cost-effective solution. Phytomining, as a new technique for extracting metals from low–grade ore or sulphide tailings, is a promising new technique currently being developed for commercialization in the United States. Field experiments in phytomining in Manitoba could provide valuable information regarding the potential of this method to remediate sites contaminated with heavy metals.

GEOLOGICAL SETTING OF THE CENTRAL MANITOBA (AU) DEPOSIT

Tailings associated with the Central Manitoba gold deposit were selected for initial phytoremediation and phytomining studies. The deposit occurs within the Archean Rice Lake greenstone belt, in the Uchi Subprovince of the Superior Province in southeastern Manitoba (Fig. GS-28-1). The belt is flanked to the north by the North Caribou Terrane (Marr, 1971; Weber, 1971; Poulsen et al., 1996) and, to the south, is transitional with the English River gneissic belt (McRitchie and Weber, 1971; Weber, 1971; Poulsen et al., 1996). The Rice Lake belt is fault bounded on the north by the Wanipigow Fault and on the south by the Manigotagan Fault.

Host rocks to the deposit belong to the Bidou Lake subgroup (Poulsen et al., 1996) and comprise arkose, tuff and chert of the Dove Lake Formation. These sedimentary rocks have been intruded by gabbro sills. Gold-bearing quartz veins at the deposit are situated within en échelon shear zones at or close to the contact between the Dove Lake sedimentary rocks and an east-southeast-trending gabbro sill (Stockwell and Lord, 1939). Five veins contributed the bulk of production at the deposit. These were the Kitchener, Eclipse, No.1 Branch, Tene 6 and Hope veins. The quartz veins were mineralized with chalcopyrite, pyrite, pyrrhotite and free gold. Between 1928 and 1938, a total of 347 801 t of ore was milled and 4287 kg of gold produced.

GEOCHEMICAL CHARACTERIZATION OF TAILINGS

Geochemical analyses of 20 tailings samples collected from three hand-augered profiles, each approximately 1 m in depth, are presented in Tables GS-28-1, -2 and -3. The data in Tables GS-28-1 and -2 are total

Table GS-28-1: Instrumental neutron activation analysis (INAA) of tailings samples (-60 mesh) from three 1 m deep, hand-augered profiles. Negative values indicate less than the lower limit of determination.

Sample ID	Au	Ag	As	Ba	Br	Ca	Co	Cr	Cs	Fe	Hf	Hg	Ir	Mo	Na	Ni	Rb	Sb	Sc
	(ppb)	(ppm)	(ppm)	(ppm)	(ppm)	(%)	(ppm)	(ppm)	(ppm)	(%)	(ppm)	(ppm)	(ppb)	(ppm)	(%)	(ppm)	(ppm)	(ppm)	(ppm)
SITE 1: 0-15 cm	3430	-5	19	220	-1	-0.5	7	100	-0.5	5.49	1.5	2	-5	-5	0.58	-100	-20	0.9	12.2
SITE 1: 15-30 cm	1210	-5	16	180	-1	0.8	47	49	-0.5	3.6	1.7	-1	-5	-5	0.74	-100	-20	0.6	8
SITE 1: 30-45 cm	795	-5	10	230	-1	0.6	25	59	-0.5	3.47	1.8	-1	-5	-5	0.75	-100	-20	0.4	8.8
SITE 1: 45-60 cm	1040	-5	7	170	-1	0.9	34	32	-0.5	3.63	1.2	-1	-5	-5	0.46	-100	-20	0.3	8.2
SITE 1: 60-75 cm	352	-5	4	120	-1	-0.5	19	37	-0.5	3.11	1	-1	-5	-5	0.41	105	-20	0.4	7.9
SITE 1: 75-90 cm	727	-5	6	260	-1	1.2	28	47	-0.5	3.33	1.5	-1	-5	-5	0.56	-100	-20	0.4	9.1
SITE 1: 90-105 cm	780	-5	6	300	-1	1	25	14	-0.5	2.57	2	-1	-5	-5	0.43	-100	22	0.3	3.7
SITE 2: 0-15 cm	2290	-5	11	130	-1	1.1	8	75	-0.5	3.82	1	1	-5	-5	0.57	-100	-20	0.3	9.1
SITE 2: 15-30 cm	1530	-5	11	150	-1	0.8	16	110	-0.5	4.66	0.7	4	-5	-5	0.51	-100	-20	0.4	12.3
SITE 2: 30-45 cm	1490	-5	11	-100	-1	-0.5	22	84	-0.5	4.08	1	3	-5	-5	0.5	-100	-20	0.3	11.1
SITE 2: 45-60 cm	1810	-5	10	-100	-1	0.5	45	99	-0.5	4.32	0.7	2	-5	-5	0.39	-100	-20	-0.2	13.1
SITE 2: 60-75 cm	2640	-5	9	-100	-1	0.9	76	90	-0.5	4.04	0.9	-1	-5	-5	0.5	-100	-20	0.4	10.2
SITE 2: 75-90 cm	2390	-5	14	220	-1	1.5	83	85	-0.5	3.88	1.4	-1	-5	-5	0.63	-100	-20	0.4	9.5
SITE 2: 90-105 cm	1110	-5	20	220	-1	1.2	54	42	-0.5	3.4	2.8	-1	-5	-5	0.99	-100	-20	0.4	6.8
SITE 3: 0-15 cm	428	-5	6	190	-1	1.6	32	99	-0.5	4.06	1.8	-1	-5	-5	0.43	-100	32	0.3	10.6
SITE 3: 15-30 cm	569	-5	-2	190	-1	1.2	36	210	-0.5	4.92	0.5	-1	-5	-5	0.27	-100	-20	0.3	15.4
SITE 3: 30-45 cm	265	-5	4	-100	-1	1.5	29	178	1.5	4.35	-0.5	-1	-5	-5	0.21	-100	21	0.6	12.8
SITE 3: 45-60 cm	378	-5	3	-100	-1	0.7	29	194	-0.5	4.58	-0.5	-1	-5	-5	0.21	-100	-20	0.6	12.7
SITE 3: 60-75 cm	578	-5	4	-100	-1	1.3	31	260	-0.5	4.98	0.5	-1	-5	-5	0.25	-100	-20	0.4	15.6
SITE 3: 75-90 cm	958	-5	5	-100	-1	0.7	35	137	-0.5	4.79	0.7	-1	-5	-5	0.48	-100	-20	0.3	14.5

Sample ID	Se	Sr	Ta	Th	U	W	Zn	La	Ce	Nd	Sm	Eu	Tb	Yb	Lu	Mass
	(ppm)	(%)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)	(g)
SITE 1: 0-15 cm	6	-0.05	-1	0.5	-0.5	6	40	1.5	4	-5	0.5	0.2	-0.5	0.8	0.11	1.485
SITE 1: 15-30 cm	3	-0.05	-1	1.1	-0.5	4	217	5.1	11	5	1.3	0.3	-0.5	0.8	0.13	1.803
SITE 1: 30-45 cm	-3	-0.05	-1	1.4	-0.5	4	110	3.7	8	-5	1.1	0.2	-0.5	0.9	0.14	1.484
SITE 1: 45-60 cm	-3	-0.05	-1	0.8	-0.5	4	187	3.5	8	-5	1	0.2	-0.5	0.8	0.12	1.597
SITE 1: 60-75 cm	-3	-0.05	-1	0.6	-0.5	-3	85	2.3	4	-5	0.7	0.2	-0.5	0.6	0.09	1.676
SITE 1: 75-90 cm	4	-0.05	-1	0.9	-0.5	4	128	3.3	8	-5	1	0.2	-0.5	0.8	0.14	1.435
SITE 1: 90-105 cm	-3	-0.05	-1	1	-0.5	5	79	5	11	5	1.2	0.2	-0.5	0.8	0.14	1.453
SITE 2: 0-15 cm	6	-0.05	-1	0.5	-0.5	5	65	1.5	4	-5	0.5	-0.1	-0.5	0.7	0.11	1.604
SITE 2: 15-30 cm	5	-0.05	-1	-0.5	-0.5	4	62	1.5	3	-5	0.5	0.2	-0.5	0.6	0.09	1.657
SITE 2: 30-45 cm	5	-0.05	-1	0.5	-0.5	4	69	1.2	3	-5	0.5	0.1	-0.5	0.6	0.1	1.57
SITE 2: 45-60 cm	5	-0.05	-1	-0.5	-0.5	3	103	1.6	4	-5	0.9	0.1	-0.5	0.8	0.12	1.605
SITE 2: 60-75 cm	6	-0.05	-1	0.6	-0.5	5	253	4.4	8	-5	1.1	0.2	-0.5	1	0.15	1.577
SITE 2: 75-90 cm	4	-0.05	-1	0.9	-0.5	4	259	4.8	9	-5	1.1	0.4	-0.5	0.9	0.15	1.362
SITE 2: 90-105 cm	-3	-0.05	-1	1.8	-0.5	6	234	7.5	15	7	1.7	0.5	-0.5	1.1	0.16	1.573
SITE 3: 0-15 cm	3	-0.05	-1	1	-0.5	4	117	3.9	9	-5	1.1	0.3	-0.5	0.9	0.14	1.833
SITE 3: 15-30 cm	5	-0.05	-1	-0.5	-0.5	3	156	1.6	5	-5	0.6	-0.1	-0.5	0.6	0.1	1.427
SITE 3: 30-45 cm	-3	-0.05	-1	-0.5	-0.5	3	75	0.9	-3	-5	0.4	-0.1	-0.5	0.5	0.07	1.372
SITE 3: 45-60 cm	-3	-0.05	-1	-0.5	-0.5	3	103	0.9	-3	-5	0.4	-0.1	-0.5	0.5	0.07	1.734
SITE 3: 60-75 cm	4	-0.05	-1	-0.5	-0.5	3	141	0.9	-3	-5	0.4	-0.1	-0.5	0.5	0.08	1.346
SITE 3: 75-90 cm	6	-0.05	-1	-0.5	-0.5	3	133	1.4	-3	-5	0.6	0.2	-0.5	0.7	0.11	1.532

Table GS-28-2: Geochemical analysis of tailings samples (-60 mesh) collected from three 1 m deep, hand-augered profiles. Analysis by ICP-MS following a lithium metaborate/tetraborate fusion. All values in ppm. Negative values indicate less than the lower limit of determination.

Sample ID	V	Cr	Co	Ni	Cu	Zn	Ga	Ge	As	Rb	Sr	Y	Zr	Nb	Mo	Ag	In	Sn	Sb	Cs	Ba
SITE 1: 0-15 cm	68	71	6	-20	834	153	8	-1	9	18	16	6	51	2	-2	6.0	-0.2	1	-0.5	-0.5	155
SITE 1: 15-30 cm	46	48	25	30	3640	82	9	1	10	20	23	8	70	2	-2	3.4	-0.2	1	0.5	-0.5	199
SITE 1: 30-45 cm	47	37	37	51	5740	183	9	-1	11	19	24	9	67	2	-2	4.0	-0.2	-1	-0.5	-0.5	177
SITE 1: 45-60 cm	40	25	31	36	3990	155	7	-1	-5	16	15	7	48	2	-2	3.9	-0.2	-1	-0.5	-0.5	150
SITE 1: 60-75 cm	42	30	17	24	3470	82	6	-1	-5	12	12	6	38	2	-2	2.8	-0.2	-1	-0.5	-0.5	120
SITE 1: 75-90 cm	46	32	26	29	3800	114	8	-1	5	19	18	8	59	2	-2	3.3	-0.2	-1	-0.5	0.5	198
SITE 1: 90-105 cm	23	-20	22	-20	3770	71	7	-1	10	19	17	7	72	2	-2	3.2	-0.2	-1	-0.5	-0.5	243
SITE 2: 0-15 cm	58	34	8	-20	1710	-30	7	-1	-5	16	18	6	46	2	-2	4.7	-0.2	-1	-0.5	-0.5	126
SITE 2: 15-30 cm	76	98	15	23	2890	43	7	-1	-5	13	13	7	30	1	-2	5.8	-0.2	-1	-0.5	-0.5	106
SITE 2: 30-45 cm	64	72	21	26	2370	44	6	-1	11	12	12	6	32	2	-2	5.2	-0.2	-1	-0.5	-0.5	106
SITE 2: 45-60 cm	64	81	35	53	10500	87	6	-1	9	11	12	6	26	1	-2	5.9	-0.2	-1	-0.5	-0.5	91
SITE 2: 60-75 cm	51	69	62	70	6560	178	7	-1	-5	15	19	10	43	2	-2	5.1	-0.2	-1	-0.5	-0.5	127
SITE 2: 75-90 cm	49	60	72	73	5860	211	8	-1	6	16	25	9	56	2	-2	4.8	-0.2	-1	-0.5	-0.5	149
SITE 2: 90-105 cm	37	29	51	44	6030	186	10	-1	9	23	37	11	101	3	-2	5.1	-0.2	-1	-0.5	-0.5	247
SITE 3: 0-15 cm	59	79	31	43	4510	111	9	-1	-5	22	21	9	65	2	-2	3.8	-0.2	-1	-0.5	-0.5	215
SITE 3: 15-30 cm	75	159	34	56	3490	93	7	-1	-5	17	10	6	24	1	-2	4.3	-0.2	-1	-0.5	-0.5	111
SITE 3: 30-45 cm	62	142	26	53	3150	68	5	-1	-5	10	8	5	13	-1	-2	4.2	-0.2	-1	-0.5	-0.5	69
SITE 3: 45-60 cm	58	141	24	50	3830	70	5	-1	-5	9	6	5	10	-1	-2	5.0	-0.2	-1	-0.5	-0.5	64
SITE 3: 60-75 cm	77	145	28	48	3510	78	6	-1	-5	15	7	5	13	-1	-2	4.7	-0.2	-1	-0.5	-0.5	89
SITE 3: 75-90 cm	72	114	30	43	4440	94	7	-1	-5	14	11	6	26	1	-2	4.8	-0.2	-1	-0.5	-0.5	121
Standards																					
Blank	-5	-20	-1	-20	-10	-30	-1	-1	-5	-2	-2	-1	-5	-1	-2	-0.5	-0.2	-1	-0.5	-0.5	-3
Standard MAG1	132	96	22	44	27	117	22	2	11	153	138	28	115	14	-2	-0.5	-0.2	3	0.9	8.5	482
Certified MAG1	**140***	**97***	**20.4***	**53***	**30***	**130***	**20.4***		**9.2**	**149***	**146***	**28***	**126***	**12**	**1.6**	**0.08**	**(0.18)**	**3.6**	**0.96***	**8.6***	**479***
Standard BIR1	310	392	51	163	125	76	16	2	-5	-2	107	16	14	-1	-2	-0.5	-0.2	-1	0.6	-0.5	7
Certified BIR1	**313***	**382***	**51.4***	**166***	**126***	**71***	**16**	**1.5**	**(0.4)**	**0.25***	**108***	**16***	**16**	**0.6**	**(0.5)**	**(0.036)**		**0.65**	**0.58**	**0.005**	**7**
Standard DNC1	139	275	55	262	99	60	15	1	-5	3	138	18	37	2	-2	-0.5	-0.2	1	0.9	-0.5	113
Certified DNC1	**148***	**285***	**54.7***	**247***	**96***	**66***	**15**	**(1.3)**	**(0.2)**	**(4.5)**	**145***	**18***	**41***	**3**	**(0.7)**	**(0.027)**			**0.96***	**(0.34)**	**114***
Standard GXR-2	49	34	7	-20	68	563	37	1	24	74	152	17	244	9	-2	17.0	-0.2	2	45.9	5.0	2300
Certified GXR-2	**52**	**36**	**8.6**	**21**	**76**	**530**	**37**		**25**	**78.0**	**160**	**17**	**269**	**11**	**(2.1)**	**17**	**(0.252)**	**1.7**	**49**	**5.2**	**2240**
Standard LKSD-3	75	85	28	45	31	138	15	1	26	76	246	30	162	8	-2	-0.5	-0.2	2	1.1	2.2	656
Certified LKSD-3	**82**	**87**	**30**	**47**	**35**	**152**			**27**	**78**	**240**	**30**	**178**	**8**	**(<5)**	**2.7**		**3**	**1.3**	**2.3**	**680**
Standard MICA-Fe	130	82	25	-20	-10	1300	97	3	3	2170	4	47	815	270	-2	-0.5	0.6	70	-0.5	179	145
Certified Mica Fe	**135***	**90***	**23***	**35***	**5***	**1300***	**95***	**3.2**	**3**	**2200***	**5***	**48***	**800***	**270***	**1.2**		**0.60**	**70***		**180***	**150***
Standard GXR1	89	-40	9	44	1120	792	16	3	443	11	287	33	30	-2	18	31	0.8	53	122	3	772
Certified GXR1	**80**	**12**	**8.2**	**41**	**1110**	**760**	**14**		**427**	**(14)**	**275**	**32**	**(38)**	**(0.8)**	**18**	**31**	**0.8**	**54**	**122**	**3.0**	**750**
Standard SY3	49	-40	7	-40	-20	265	27	-2	20	218	308	723	302	164	-4	-1	-0.4	8	-1	3	449
Certified SY3	**50**	**(11)**	**8.8**	**11**	**17**	**244***	**27***	**1.4**	**19**	**206***	**302***	**718***	**320**	**148**	**(1.0)**	**(1.5)**		**(6.5)**	**0.31**	**3**	**450**
Standard STM-1	-5	-20	-1	-20	-10	248	36	2	-5	112	695	45	1210	247	5	-0.5	-0.2	7	1.4	1.5	562
Certified STM-1	**(8.7)**	**(4.3)**	**0.9**	**(3)**	**(4.6)**	**235***	**36***	**(1.4)**	**4.6**	**118***	**700***	**46***	**1210***	**268***	**5.2**	**0.079***	**(0.12)**	**6.8**	**1.66***	**1.54***	**560***
Standard IFG-1	-5	-20	31	-20	13	-30	1	23	-5	-2	5	11	-5	-1	-2	-0.5	-0.2	-1	0.8	-0.5	7
Certified IFG-1	**2**	**4**	**29***	**22.5**	**13***	**20***	**0.7**	**24**	**1 5**	**0.4**	**3**	**9***	**1**	**0.1***	**0.7**		**0.2**	**0.3**	**0.63**	**0.06**	**1.5**

Table GS-28-2: Geochemical analysis of tailings samples (-60 mesh) collected from three 1 m deep, hand-augered profiles. Analysis by ICP-MS following a lithium metaborate/tetraborate fusion. All values in ppm. Negative values indicate less than the lower limit of determination. ***(continued)***

Sample ID	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Hf	Ta	W	Tl	Pb	Bi	Th	U
SITE 1: 0-15 cm	1.7	3.1	0.42	1.6	0.4	0.12	0.5	0.1	0.9	0.2	0.8	0.12	0.8	0.13	1.3	0.2	7	-0.1	33	127	0.8	0.2
SITE 1: 15-30 cm	3.4	7.5	0.89	3.9	1.0	0.25	1.1	0.2	1.3	0.3	0.9	0.14	0.8	0.13	1.8	0.2	7	-0.1	36	46.0	1.6	0.3
SITE 1: 30-45 cm	4.4	9.3	1.11	5.0	1.1	0.27	1.3	0.2	1.4	0.3	0.9	0.14	0.9	0.13	1.7	0.2	7	0.1	33	53.0	1.3	0.3
SITE 1: 45-60 cm	3.0	6.7	0.80	3.6	0.9	0.21	1.0	0.2	1.0	0.2	0.7	0.11	0.7	0.11	1.2	0.2	5	-0.1	18	42.0	1.0	0.2
SITE 1: 60-75 cm	2.0	4.5	0.56	2.5	0.6	0.16	0.7	0.1	0.9	0.2	0.6	0.10	0.6	0.09	1.0	0.1	3	-0.1	19	49.0	0.6	0.1
SITE 1: 75-90 cm	3.2	7.0	0.85	3.8	0.9	0.23	1.0	0.2	1.2	0.3	0.8	0.13	0.9	0.13	1.6	0.2	4	-0.1	31	44.0	1.0	0.2
SITE 1: 90-105 cm	4.3	9.3	1.10	4.6	1.0	0.24	1.1	0.2	1.1	0.2	0.7	0.12	0.7	0.12	1.9	0.2	9	-0.1	23	42.0	1.3	0.2
SITE 2: 0-15 cm	1.7	3.3	0.44	1.8	0.4	0.11	0.6	0.1	0.9	0.2	0.7	0.11	0.7	0.11	1.2	0.1	8	-0.1	16	61.0	0.5	0.1
SITE 2: 15-30 cm	1.7	3.2	0.44	1.9	0.5	0.12	0.7	0.2	1.0	0.2	0.8	0.12	0.7	0.11	0.8	-0.1	6	-0.1	30	92.0	0.5	0.1
SITE 2: 30-45 cm	1.4	2.9	0.41	1.8	0.5	0.12	0.7	0.1	0.9	0.2	0.7	0.11	0.7	0.10	0.9	0.1	5	-0.1	28	89.0	0.6	0.1
SITE 2: 45-60 cm	1.4	3.8	0.53	2.7	0.8	0.19	0.9	0.2	1.0	0.2	0.7	0.11	0.7	0.11	0.7	-0.1	4	-0.1	41	98.0	0.4	0.3
SITE 2: 60-75 cm	4.0	8.2	1.02	4.5	1.1	0.31	1.5	0.3	1.5	0.3	1.0	0.15	0.9	0.14	1.1	0.1	5	-0.1	18	94.0	0.7	0.2
SITE 2: 75-90 cm	4.3	9.0	1.08	4.8	1.1	0.29	1.4	0.2	1.4	0.3	0.9	0.14	0.8	0.14	1.5	0.2	6	-0.1	20	78.0	1.0	0.2
SITE 2: 90-105 cm	6.8	14.9	1.80	7.8	1.6	0.42	1.9	0.3	1.8	0.4	1.2	0.17	1.1	0.17	2.6	0.3	7	-0.1	31	41.0	1.8	0.4
SITE 3: 0-15 cm	3.7	8.1	1.03	4.4	1.0	0.26	1.2	0.2	1.4	0.3	0.9	0.15	0.9	0.14	1.7	0.2	6	-0.1	88	66.0	1.1	0.2
SITE 3: 15-30 cm	1.6	3.2	0.46	2.0	0.6	0.18	0.8	0.2	1.0	0.2	0.7	0.10	0.7	0.11	0.7	-0.1	4	-0.1	70	79.0	0.3	0.1
SITE 3: 30-45 cm	1.0	1.9	0.29	1.3	0.4	0.13	0.6	0.1	0.7	0.2	0.5	0.08	0.5	0.08	0.4	-0.1	4	-0.1	51	78.0	0.1	-0.1
SITE 3: 45-60 cm	1.1	1.9	0.29	1.3	0.4	0.11	0.5	0.1	0.8	0.2	0.5	0.08	0.5	0.07	0.3	-0.1	3	-0.1	51	89.0	0.1	-0.1
SITE 3: 60-75 cm	0.8	2.0	0.28	1.3	0.4	0.12	0.6	0.1	0.9	0.2	0.6	0.10	0.6	0.09	0.4	-0.1	3	-0.1	58	87.0	0.2	-0.1
SITE 3: 75-90 cm	1.3	3.0	0.39	1.9	0.5	0.13	0.7	0.1	0.9	0.2	0.6	0.11	0.7	0.11	0.7	-0.1	5	-0.1	43	62.0	0.3	-0.1
Standards																						
Blank	-0.1	-0.1	-0.05	-0.1	-0.1	-0.05	-0.1	-0.1	-0.1	-0.1	-0.1	-0.05	-0.1	-0.04	-0.2	-0.1	-1	-0.1	-5	-0.4	-0.1	-0.1
Standard MAG1	43.9	89.4	9.84	38.1	7.3	1.52	6.6	1.0	5.3	1.0	2.9	0.44	2.7	0.38	3.5	1.2	2	0.3	21	-0.4	12.1	2.9
Certified MAG1	**43***	**88***	**9.3**	**38***	**7.5***	**1.55***	**5.8***	**0.96***	**5.2***	**1.02***	**3**	**0.43***	**2.6***	**0.40***	**3.7***	**1.1**	**1.4**	**(0.59)**	**24***	**0.34**	**11.9***	**2.7***
Standard BIR1	0.8	2.1	0.40	2.5	1.1	0.56	1.8	0.4	2.6	0.6	1.8	0.28	1.7	0.28	0.6	-0.1	-1	-0.1	-5	-0.4	-0.1	-0.1
Certified BIR1	**0.62***	**1.95***	**0.38***	**2.5***	**1.1***	**0.54***	**1.85***	**0.36***	**2.5***	**0.57***	**1.7***	**0.26***	**1.65**	**0.26***	**0.6***	**0.04**	**0.07**	**(0.01)**	**3**	**(0.02)**	**0.03**	**0.01**
Standard DNC1	3.7	10.1	1.04	4.7	1.4	0.61	2.0	0.4	2.8	0.6	2.0	0.32	2.0	0.32	1.0	-0.1	-1	-0.1	6	-0.4	0.3	-0.1
Certified DNC1	**3.8***	**10.6**	**1.3**	**4.9***	**1.38***	**0.59***	**2**	**0.41***	**2.7**	**0.62**	**2***	**(0.33)**	**2.01***	**0.32***	**1.01***	**0.098***	**(0.2)**	**(0.026)**	**6.3**	**(0.02)**	**(0.2)**	**(0.1)**
Standard GXR-2	23.9	48.4	4.87	18.3	3.4	0.69	3.1	0.5	2.7	0.6	1.7	0.27	1.7	0.26	6.4	0.8	1	1.0	676	-0.4	8.1	2.7
Certified GXR-2	**25.6**	**51.4**		**(19)**	**3.5**	**0.81**	**(3.3)**	**0.48**	**3.3**			**(0.3)**	**2.04**	**(0.27)**	**8.3**	**0.9**	**1.9**	**1.03**	**690**	**(0.69)**	**8.8**	**2.9**
Standard LKSD-3	48.8	91.6	11.2	43.2	7.7	1.51	6.7	0.9	5.1	1.0	3.0	0.46	2.8	0.41	4.6	0.6	-1	0.2	27	-0.4	11.3	4.5
Certified LKSD-3	**52**	**90**		**44**	**8.0**	**1.50**		**1.0**	**4.9**				**2.7**	**0.4**	**4.8**	**0.7**	**(<4)**		**29**		**11.4**	**4.6**
Standard MICA-Fe	199	412	48.6	179	32.9	0.70	24.9	2.7	10.6	1.4	3.4	0.48	2.9	0.25	26.1	34.4	16	17.4	14	1.7	166	85.3
Certified Mica Fe	**200***	**420***	**49***	**180***	**33***	**0.7***	**21***	**2.7***	**11***	**1.6***	**3.8***	**0.48***	**3.5***	**0.5***	**26***	**35***	**15**	**16**	**13***	**2**	**150***	**80***
Standard GXR1	9.0	17	2.2	9.6	3.1	0.7	4.4	0.9	4.6	1.1	3.2	0.5	2.2	0.32	0.8	-0.2	168	0.5	702	1380	3.0	34.7
Certified GXR1	**7.5**	**17**		**(18)**	**2.7**	**0.69**	**4.2**	**0.83**	**4.3**			**(0.43)**	**1.9**	**0.3**	**1.0**	**0.175**	**164**	**(0.39)**	**730**	**1380**	**2.44**	**34.9**
Standard SY3	1340	2220	223	671	110	17.2	105	18.2	118	29.6	68.4	11.6	62.2	7.90	12.4	27.2	2	1.6	133	2.9	1000	650
Certified SY3	**1340***	**2230***	**223***	**670**	**109**	**17***	**105***	**18**	**118**	**29.5***	**68**	**11.6***	**(62)**	**7.90**	**9.70**	**30***	**1.1***	**1.50**	**133***	**(0.8)**	**1003***	**650***
Standard STM-1	147	256	20.9	76.3	11.7	3.54	10.1	1.5	8.0	1.5	4.4	0.67	4.3	0.58	27.6	19.0	4	0.2	18	1.5	31.4	9.0
Certified STM-1	**150***	**259***	**19***	**79***	**12.6***	**3.6***	**9.5***	**1.55***	**8.1***	**1.9**	**4.2***	**0.69**	**4.4***	**0.60**	**28***	**18.6***	**3.6***	**0.26**	**17.7***	**0.13**	**31***	**9.06***
Standard IFG-1	3.3	4.8	0.52	2.1	0.5	0.41	0.8	0.1	1.0	0.2	0.8	0.11	0.7	0.10	-0.2	0.2	219	-0.1	-5	-0.4	0.2	0.2
Certified IFG-1	**2.8***	**4***	**0.4***	**0.2**	**0.4***	**0.39***	**0.74***	**0.11***	**0.8***	**0.2***	**0.63***	**0.09***	**0.6***	**0.09***	**0.04**	**0.2**	**220**	**0.02**	**4**		**0.1**	**0.02**

*Recommended value.
( ) Information value.
All other values are proposed.

Table GS-28-3: Geochemical analysis of tailings samples (-60 mesh) collected from three 1 m deep, hand-augered profiles. Analysis by ICP-MS following an ammonium iodide extraction. Negative values indicate less than the lower limit of determination.

Sample ID	Li*	Be*	Cl*	Sc*	Ti*	V	Mn	Co	Ni	Cu	Zn	Ga	Ge	As	Se	Br	Rb	Sr	Y	Zr	Nb	Mo	Ru
SITE 1: 0-15 cm	6	-2	48384	-100	939	102	643	69	100	65055	126	12	2	40	25	6349	27	346	34	131	-1	7	1
SITE 1: 15-30 cm	396	4	46760	-100	460	60	50541	5529	2072	1691987	6479	21	5	127	45	6598	17	681	677	104	-1	25	-1
SITE 1: 30-45cm	32	-2	42286	-100	276	64	113460	10392	3373	1589745	34229	19	4	474	47	6528	19	1465	662	107	-1	51	-1
SITE 1: 45-60 cm	41	-2	46835	-100	244	38	58022	4647	1695	704168	24420	10	3	171	37	7387	14	1178	200	132	-1	32	-1
SITE 1: 60-75 cm	15	-2	47159	-100	195	42	33047	729	882	735417	5850	11	6	165	39	7388	8	1779	220	136	-1	22	-1
SITE 1: 75-90 cm	13	-2	47853	-100	516	79	50233	1796	1811	561936	8323	22	7	257	62	7293	11	2233	303	243	1	29	-1
SITE 1: 90-105 cm	7	-2	49338	-100	570	78	66712	390	799	149146	4260	31	11	188	54	7707	14	1837	220	220	1	24	-1
SITE 2: 0-15 cm	122	-2	53883	116	1995	174	12194	1026	1044	358362	1775	13	1	60	33	8066	133	904	133	130	2	6	1
SITE 2: 15-30 cm	314	4	52059	206	1603	435	48501	6717	3404	767980	4914	17	3	107	58	7977	189	919	311	110	1	9	-1
SITE 2: 30-45 cm	204	5	50391	228	369	276	55017	7087	3397	475237	4011	11	3	140	73	7580	110	535	176	98	-1	7	-1
SITE 2: 45-60 cm	232	24	45438	621	400	513	31700	5194	3337	999999	3681	19	9	140	92	6188	67	741	335	66	-1	9	-1
SITE 2: 60-75 cm	43	5	50442	147	505	401	257171	15623	6502	587436	35676	40	9	186	97	8639	64	2026	2082	180	1	57	-1
SITE 2: 75-90 cm	2082	4	52355	107	2952	288	185847	17540	7625	328337	39056	33	10	148	68	7988	48	1963	1090	172	2	73	-1
SITE 2: 90-105 cm	86	4	48230	108	950	164	154497	7183	3336	262740	32696	45	15	125	44	8553	22	2255	810	416	2	103	3
SITE 3: 0-15 cm	4	-2	43766	-100	264	99	64393	1661	1565	724289	5703	15	4	208	83	7902	32	1305	291	226	-1	26	-1
SITE 3: 15-30 cm	-2	2	44070	143	157	180	61800	1714	1499	700695	4877	11	3	98	121	8425	102	2057	424	92	-1	22	-1
SITE 3: 30-45 cm	3	-2	42244	119	126	155	47503	741	1155	613241	4052	9	2	55	70	8189	13	1835	351	39	-1	17	2
SITE 3: 45-60 cm	273	-2	40126	-100	154	172	36456	688	1003	647383	3627	9	3	62	67	7137	16	1389	294	47	-1	16	-1
SITE 3: 60-75 cm	6	-2	41281	103	171	262	38533	1127	1600	626482	4051	10	3	57	78	7499	22	1538	343	63	-1	9	1
SITE 3: 75-90 cm	10	-2	40266	-100	257	210	44879	1081	1113	323513	3544	13	4	65	63	7518	47	1729	307	87	-1	10	-1

Sample ID	Pd	Ag	Cd	In	Sn	Sb	Te	Cs	Ba	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Hf	Ta
SITE 1: 0-15 cm	1	492.2	7.0	2.7	13	2.9	15	2.1	-1	79.7	200.2	25.9	109.7	19.5	3.2	17.8	3.0	10.0	1.6	4.9	1.1	3.3	0.5	3.6	-0.1
SITE 1: 15-30 cm	1	276.7	129.7	5.7	12	5.4	90	1.7	168	599.4	1180.8	153.1	666.0	140.7	32.5	141.3	27.3	133.2	25.5	68.3	7.9	55.4	8.9	5.7	0.2
SITE 1: 30-45cm	2	133.2	339.3	7.5	22	8.5	295	1.2	280	471.7	857.4	107.4	453.5	98.3	25.9	109.2	21.6	112.8	22.7	63.3	8.6	53.4	9.4	5.0	0.3
SITE 1: 45-60 cm	3	107.6	262.5	9.9	18	4.6	215	0.6	271	117.7	236.1	28.8	126.6	29.7	7.5	31.9	6.0	34.4	6.9	20.2	2.7	19.0	4.1	5.1	-0.1
SITE 1: 60-75 cm	2	128.6	329.1	11.3	28	10.7	332	0.4	218	120.2	238.7	29.5	135.8	30.6	7.7	33.1	6.6	37.8	7.6	22.6	3.4	23.5	4.8	4.8	-0.1
SITE 1: 75-90 cm	3	152.2	311.9	12.5	44	29.3	649	0.8	296	214.2	445.7	52.9	246.7	55.3	13.9	55.7	10.2	55.6	10.9	32.1	3.9	30.1	6.0	8.3	-0.1
SITE 1: 90-105 cm	3	157.4	85.3	5.9	30	43.6	1653	0.5	516	232.8	481.8	56.0	238.8	51.9	13.2	50.2	9.1	45.0	8.2	23.6	2.7	19.5	3.1	6.2	-0.1
SITE 2: 0-15 cm	1	295.1	62.9	5.7	15	3.0	42	9.7	-1	132.9	332.8	43.1	185.3	33.3	6.8	33.1	6.6	27.5	5.3	15.7	1.7	11.1	1.7	3.3	-0.1
SITE 2: 15-30 cm	1	339.7	208.4	23.7	14	7.4	130	29.5	3	176.9	477.6	60.9	290.0	67.0	15.2	66.6	13.2	67.2	13.1	36.6	4.5	28.8	4.4	3.2	-0.1
SITE 2: 30-45 cm	-1	374.6	156.2	30.5	27	9.0	288	26.2	39	121.9	307.8	42.7	215.5	49.7	10.8	43.9	8.3	38.3	7.4	22.5	2.8	19.4	3.0	2.8	-0.1
SITE 2: 45-60 cm	-1	153.4	111.5	7.5	20	18.8	727	10.8	32	229.7	798.6	129.9	678.5	202.8	44.3	130.5	26.6	132.7	24.2	71.7	10.2	83.1	12.9	3.0	0.1
SITE 2: 60-75 cm	-1	157.0	1919.0	4.7	96	39.7	416	5.1	239	1044.4	2005.7	236.4	1099.0	274.4	79.3	327.7	69.2	372.7	73.5	208.1	26.7	178.7	29.7	8.4	0.5
SITE 2: 75-90 cm	-1	150.4	1282.1	3.0	43	38.3	313	4.5	274	758.9	1472.4	169.8	737.4	158.2	44.2	198.1	37.5	191.6	37.9	104.0	12.7	88.0	14.9	6.8	0.2
SITE 2: 90-105 cm	3	155.9	940.8	3.0	53	39.3	211	2.1	436	842.3	1882.0	221.7	897.8	151.7	46.7	219.3	39.2	184.3	33.8	95.9	11.8	80.0	13.9	14.3	0.3
SITE 3: 0-15 cm	1	137.7	228.0	11.8	44	9.3	613	2.3	517	203.8	409.0	49.9	229.1	54.0	13.5	53.9	10.6	53.6	10.9	30.3	5.0	29.5	5.6	8.1	-0.1
SITE 3: 15-30 cm	-1	157.1	479.6	10.7	18	10.4	167	2.2	387	149.2	313.2	40.9	199.2	52.5	16.7	57.3	12.3	71.2	15.6	45.8	6.7	44.4	8.5	3.6	-0.1
SITE 3: 30-45 cm	-1	181.8	386.3	10.9	40	10.5	71	1.9	230	110.2	230.0	28.8	153.7	40.8	14.4	45.9	9.9	58.3	12.6	35.4	5.3	36.1	6.9	1.4	-0.1
SITE 3: 45-60 cm	-1	213.8	259.7	13.1	19	14.1	90	1.9	217	91.3	188.9	28.4	131.3	35.7	11.5	38.9	8.2	49.2	10.5	31.6	4.2	31.0	5.8	2.1	-0.1
SITE 3: 60-75 cm	-1	236.5	297.7	10.3	51	21.4	96	2.6	273	105.9	224.3	29.1	148.2	42.2	13.7	45.3	9.7	59.9	12.3	37.5	4.9	34.9	6.6	2.5	-0.1
SITE 3: 75-90 cm	-1	234.8	197.6	6.5	32	20.0	199	5.3	517	139.7	304.9	39.7	197.4	51.3	13.4	50.4	10.2	55.3	11.3	34.5	4.0	31.8	6.1	3.6	-0.1

*Element determined semiquantitatively.
Values of 999999 indicate greater than the upper limit of detection.

Table GS-28-3: Geochemical analysis of tailings samples (-60 mesh) collected from three 1 m deep, hand-augered profiles. Analysis by ICP-MS following an ammonium iodide extraction. Negative values indicate less than the lower limit of determination. ***(continued)***

Sample ID	W	Re	Os	Pt	Au	S.Q.Hg	Tl	Pb	Bi	Th	U
SITE 1: 0-15 cm	3	-0.01	-1	-1	157.77	854	1.9	123	10018	8.8	1.8
SITE 1: 15-30 cm	7	0.07	-1	-1	129.79	328	-0.1	1808	6155	30.4	11.4
SITE 1: 30-45cm	10	0.03	-1	-1	175.99	319	3.4	1823	10612	16.0	9.5
SITE 1: 45-60 cm	9	0.04	-1	-1	125.37	128	-0.1	689	6718	11.0	7.0
SITE 1: 60-75 cm	8	0.10	-1	-1	48.96	101	-0.1	789	10690	10.1	6.4
SITE 1: 75-90 cm	21	0.34	-1	-1	43.42	107	3.7	2689	13731	28.4	9.5
SITE 1: 90-105 cm	100	0.19	-1	-1	22.69	66	6.6	4584	20620	37.1	5.3
SITE 2: 0-15 cm	6	0.14	-1	-1	163.27	702	-0.1	49	11129	9.1	2.0
SITE 2: 15-30 cm	12	0.31	-1	-1	196.75	1612	-0.1	232	22920	53.8	3.9
SITE 2: 30-45 cm	11	0.49	-1	-1	95.09	1100	-0.1	328	30381	31.3	4.3
SITE 2: 45-60 cm	17	0.49	-1	-1	35.04	216	-0.1	18972	28175	13.6	86.7
SITE 2: 60-75 cm	24	0.76	-1	-1	1.67	203	-0.1	8266	17087	18.8	23.4
SITE 2: 75-90 cm	16	0.65	-1	-1	2.60	225	-0.1	8151	9929	20.6	17.2
SITE 2: 90-105 cm	18	0.41	-1	-1	3.78	292	-0.1	13811	4502	51.2	11.0
SITE 3: 0-15 cm	26	0.06	-1	-1	48.08	351	-0.1	6878	9932	20.6	10.4
SITE 3: 15-30 cm	21	0.03	-1	-1	78.52	213	-0.1	3253	10202	5.9	14.6
SITE 3: 30-45 cm	8	0.02	-1	-1	40.65	139	-0.1	2904	12313	2.0	12.9
SITE 3: 45-60 cm	9	0.17	-1	-1	22.43	90	-0.1	2659	17144	2.1	11.8
SITE 3: 60-75 cm	8	0.25	-1	-1	6.07	131	-0.1	7068	14671	2.9	14.0
SITE 3: 75-90 cm	14	0.26	-1	-1	11.81	170	0.6	3153	9844	7.3	9.8

*Element determined semiquantitatively.
Values of 999999 indicate greater than the upper limit of detection.

analyses based on instrumental neutron activation analysis (INAA) and inductively coupled plasma mass spectrometry (ICP-MS), respectively. Data in Table GS-28-3 were determined by ICP-MS following an ammonium iodide partial extraction and are therefore partial analyses. This analytical approach was undertaken to simulate a soil–plant root micro-environment with a pH of between 2 and 3. In this way, an initial estimate of bio-available metal could be obtained.

RESULTS

The INAA total data from Tables GS-28-1 and -2 indicate the relative base- and precious-metal enrichment in tailings at the minesite. In particular, Au is elevated in the upper 15 cm of the tailings profile at site 1 (3450 ppb), is consistently enriched throughout the profile at site 2 (1110–2640 ppb), and is somewhat lower but elevated at site 3 (265–958 ppb). Exceptional concentrations of Cu are documented from all three sites (834–5740 ppm at site 1, 1710–10 500 ppm at site 2 and 3150–4510 ppm at site 3; Table GS-28-2). Additional enrichments of Ag (up to 6 ppm) and Bi (up to 127 ppm) are documented at site 1. Arsenic is low (<2–20 ppm) at all three sites (Table GS-28-1).

The relative enrichment of base and precious metals is observed throughout the profiles, although the highest single-sample responses tend to be situated in the upper 15–30 cm. Despite some upward movement of metal-enriched waters through evapotranspiration or fluctuation of groundwater levels near surface, metal enrichment may be critical, since root access of hyperaccumulator species will likely not exceed 0.5–1.0 m. Interestingly, previous studies reported in the literature (Robinson et al., 1999) demonstrate that multiple crops of nickel-hyperaccumulator species could be removed from nickeliferous soils before any reduction of nickel could be documented in the soil. This probably argues for movement of metal from the base to the top of the nickel-enriched soil profile.

Simulation of a plant root–soil micro-environment (pH of 2–3) for this orientation survey was necessary because not all metal present in the substrate will likely be available for uptake by plants (bio-available). The differences in bio-availability in the tailings may be explained, in part, by the mineralogy and forms of the metals present. These differences probably account for some of the variations in the concentrations of metals extracted from the samples using total versus partial techniques (cf. Tables GS-28-1, -2 and -3). Neutron activation and ICP-MS analysis based on lithium metaborate/tetraborate fusion of the samples will reflect total amounts of metals regardless of mineralogy, whereas ammonium iodide extraction will liberate metals at pH 2–3. The ICP-MS analyses following ammonium iodide extraction (Table GS-28-3) indicate the numerous base and precious metals that are being liberated from the three tailings profiles at the Central Manitoba site. These include significant quantities of Cu, Zn, Ni, Co, Mo, Cd, In, Sb, Hg, Pb, Bi and, in particular, Au and Ag. The metals defined as bio-available by the ammonium iodide extraction would be the focus of phytoremediation and the production of a bio-ore in the next phase of this study. Bio-availability of gold and base metals can be substantially increased by application of thiocyanate (Anderson et al., 1999) and chelating agents (Blaylock et al., 1997), respectively.

Field Study: Plant Species

Seedlings of native plant species of the boreal forest were selected for study. In July 2000, the following species were planted on three different sites at the Central Manitoba tailings site: dogwood (*Cornus stolonifera*), yellow willow (*Salix lutea*), white spruce (*Picea glauca*), jack pine (*Pinus banksiana*), tamarack (*Larix laricina*) and bog birch (*Betula glandulosa*). Half of the plants were fertilized with 10-30-15-04 (nitrogen-phosphate-potash-sulphur) at a rate of 500 kg/ha. Seeds of the following species were also planted: *Cornus stolonifera*, Indian mustard (*Brassica juncea*), slender wheatgrass (*Agropyron trachycaulum*) and altai wildrye (*Elymus angustus*). In the fall, the seedlings will be evaluated for survival and degree of injury. Selected organs of some woody seedlings will be harvested for metal-content measurements and the remainder will be left on site for further survival studies. The remainder of the live herbaceous seedlings will be treated with thiocyanate and harvested to evaluate their base-metal and gold content.

Greenhouse Study: Germination Rates

Seeds of *Brassica juncea*, *Sinapis alba* (white mustard), *Agropyron trachycaulum*, *Elymus angustus*, *Pinus banksiana* and *Picea glauca*

were planted in trays on tailings collected from the top 15 cm of the three selected sites. The trays were sprayed with distilled water regularly to keep moisture level relatively constant. Germination rates were recorded regularly. Red-osier dogwood seedlings were planted in two-litre pots containing tailings collected from the top 15 cm of the three selected sites. The pots were placed in trays so that the plants could be bottom watered to avoid leaching. Four replicates per site were used for the studies.

Preliminary results show that seeds were able to germinate in tailings collected from sites 1 and 3 (Fig. GS-28-2). Indian mustard, white mustard and altai wildrye had a relatively high germination rate when planted on tailings from sites 1 and 3, both of which have high gold and copper concentrations. Relatively few altai wildrye seeds were able to germinate on tailings collected from site 2. These preliminary results suggest that some of the selected plant species can tolerate the high level of metals in this environment (Fig. GS-28-3 and -4).

FUTURE WORK

Future work will focus on the determination of metal content in the plants and the facilitation of metal uptake through plant and substrate remedial treatment. Additional plant species will be tested for their ability to acquire and store base and precious metals in this environment. Further experiments with *Brassica juncea*, the most promising plant species for gold phytomining, should establish the most effective experimental and field conditions for maximizing gold uptake in a bio-ore based on this species. The economics of commercial production of bio-ores will be determined.

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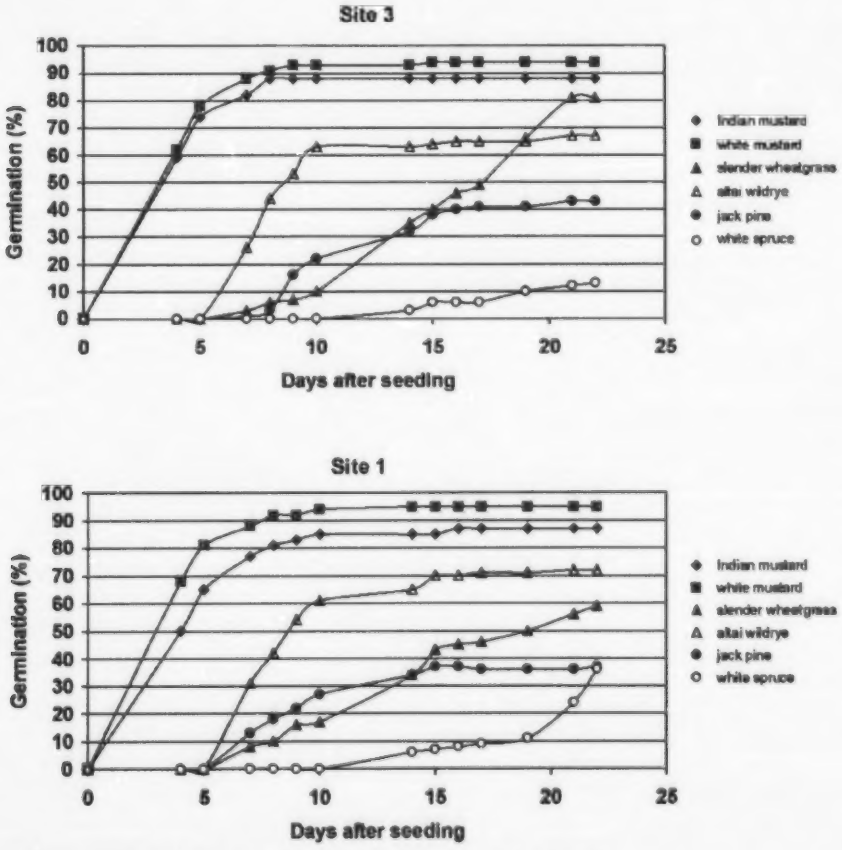

Figure GS-28-2: Germination profiles for sites 1 and 3, Central Manitoba (Au) minesite.

Figure GS-28-3: Seedlings of various species planted in July 2000 at site 2, Central Manitoba (Au) minesite.

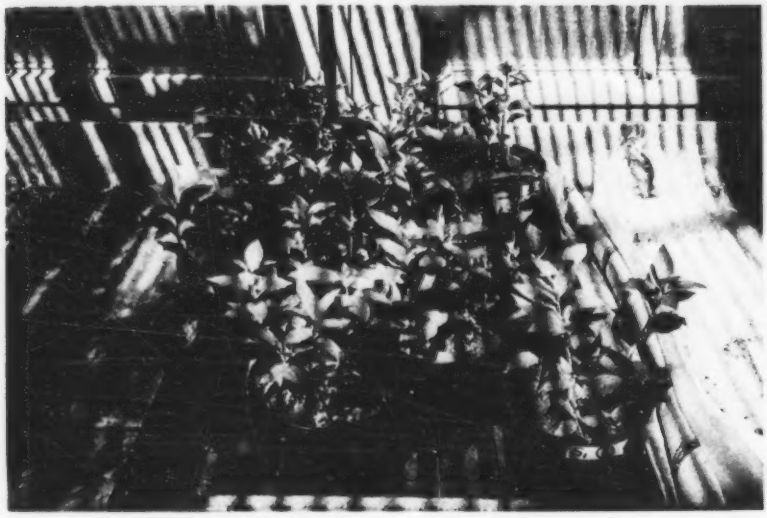

Figure GS-28-4: Red-osier dogwood seedlings growing at the University of Manitoba greenhouse in tailings from the Central Manitoba (Au) minesite.

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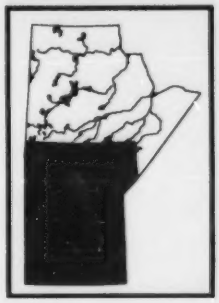
GS-29

GEOLOGY OF THE WINNIPEG REGION NATMAP PROJECT: 1999-2000 PROGRESS (NTS 62H/W, 62I AND 52L/W)



by G.L.D. Matile, L.H. Thorleifson[1], J.D. Bamburak, R.K. Bezys, A. Burt[2], G.G. Conley, N. Grant[3], G.R. Keller, J. Mann[4], C.R. McGregor and M. Pyne[1]

Matile, G.L.D., Thorleifson, L.H., Bamburak, J.D., Bezys, R.K., Burt, A., Conley, G.G., Grant, N., Keller, G.R., Mann, J., McGregor, C.R. and Pyne, M. 2000: Geology of the Winnipeg Region NATMAP Project: 1999-2000 Progress (NTS 62H/W, 62I and 52L/W); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 189-193.

SUMMARY

As a contribution to the National Geoscience Mapping Program (NATMAP), Manitoba Geological Survey (MGS) and Geological Survey of Canada (GSC) are co-operating in a program of geological mapping for the Winnipeg region, with emphasis on engineering and environmental geology. The surficial component of the NATMAP project was completed in two phases for the area from 49° to 51°N and from 95° to 98°W. In addition to completion of associated topical investigations, current emphasis is on compilation of the NATMAP outputs, and other co-operative GSC work in the area, into a synthesis. When combined with other MGS programs, such as the Capital Region Study, this synthesis will lead to generation of a three-dimensional (3-D) digital geological model for all deposits above the Precambrian, first in the NATMAP area and then for all Phanerozoic terrane in southern Manitoba. The inputs to this model include a digital elevation model, large lake bathymetry, offshore seismic surveys, surficial geology maps, lithological data for the Quaternary from water-well and other drillhole and geophysical databases, bedrock surface elevations, bedrock maps, Phanerozoic stratigraphic data and previous Phanerozoic models, and sub-Phanerozoic Precambrian geological maps. Completion of all project outputs is planned for March 2002.

INTRODUCTION

The NATMAP program was established by Canada's federal, provincial, industry, and academic geoscience communities in 1990 to promote multidisciplinary, co-operative, computer-based programs of new geological mapping. These programs would include opportunities to address mineral-resource development, environmental and societal concerns, fundamental geological knowledge, and the training of student geologists in mapping procedures.

Under the Prairie NATMAP project, initiated in 1991, new 1:100 000 scale surficial geology maps were completed for areas southeast of Winnipeg, south of latitude 50°N and east of longitude 97°W (Matile et al., 1998). Under the current Winnipeg NATMAP project (1997–2001), similar mapping and glacial sediment sampling have been completed over an expanded area, south of latitude 51°N and east of longitude 98°W; release of the maps is pending.

The objectives of this expanded, Winnipeg Region NATMAP Project are to:

1) obtain an enhanced understanding of the environmental framework and geological history of the Winnipeg region, through the synthesis of available information and collection of new field data, and to communicate this knowledge to users, primarily in the form of new, computer-based geological maps;
2) make major strides in understanding geological features such as the Belair–Sandilands glaciofluvial complex;
3) further the establishment of a Winnipeg- and Ottawa-based infrastructure for the rapid production of high-quality, interactive, digital cartographic products;
4) support the training of field geologists in the production of new maps;
5) facilitate mineral exploration, particularly in the exposed Precambrian Shield east of the Winnipeg River, by producing new geological and geochemical maps of the area;
6) provide an upgraded information base designed to support construction and other engineering activity;
7) better define geological factors that control the quantity, quality and long-term sustainability of groundwater resources in the Winnipeg region;
8) support efforts to manage the Lake Winnipeg basin, by interpreting the evolution of the lake in recent geological time and the role played by geology in controlling shoreline erosion; and
9) support environmental and land-use management, by mapping the composition and extent of lithological units that are relevant to issues such as waste disposal, soil geochemistry, and vulnerability of aquifers to contamination.

Mapping of the study area is now being extended into the subsurface, using drillhole data and geophysical surveys, in order to obtain a 3-D model. The NATMAP results, combined with other programs of the Manitoba Geological Survey, including Phanerozoic stratigraphic investigations and the Capital Region Study, are now producing a geological model with many applications, from basic science to activity such as groundwater-flow modelling and industrial-mineral resource management. Key inputs to the synthesis are also being generated by the completion of offshore and shoreline surveys in Lake Winnipeg (Thorleifson et al., 1998a), regional hydrogeological investigations (Thorleifson et al., 1998b), and studies of the Red River and its flood history (Thorleifson et al., 1998c).

This paper focuses on progress toward a digital 3-D model for all materials above the Precambrian surface, initially for the NATMAP study area (lat. 49–51°N, long. 95–98°W), and subsequently for all Phanerozoic terrane in southern Manitoba.

THREE-DIMENSIONAL MODEL

Digital Elevation Model

A detailed digital elevation model (DEM) was required 1) as the surface from which the 3-D model will hang, 2) to vertically position sites in drillhole databases, and 3) to satisfy elevation-sensitive applications such as groundwater flow. Having built the DEM, however, it was found that the model provides profound insights into both known and previously unknown geological features. Detailed data were derived from Province of Manitoba orthophoto surveys for much of southern Manitoba (Matile and Keller, 1999), and the model is now in its second version (Keller and Matile, in prep.). In order to build an elevation model for all of southern Manitoba Phanerozoic terrane, additional data from National Topographic System (NTS) contour lines and elevation points, lake-surface elevations, and United States Geological Survey (USGS) elevation data were added to the second version of the provincial data, to extend the model to latitude 55°N (Pyne et al., in prep.). In

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order to facilitate correlation of Manitoba geology with that of adjacent areas, the DEM was extended into Saskatchewan to longitude 102°W, into Ontario to longitude 94°W, and into the United States to latitude 48°N. The resulting model (Fig. GS-29-1) has a grid resolution of 100 m, and absolute vertical accuracy of about ±3 m in southern Manitoba where detailed provincial data are available. Relative accuracy is on the order of tenths of a metre, based on the scale of features defined by the model. Outside this area, absolute accuracy varies, depending on the source of the data.

Bathymetry

A full 3-D geological model and its applications (e.g. interpretation of regional bedrock structure and groundwater flow modelling) require clarification of water depths and geometry of lake-bottom features. Bathymetric information is therefore required, at least for large lakes. Soundings reported on 22 Canadian Hydrographic Service charts for Lake Winnipeg, Lake Manitoba, Lake Winnipegosis, Playgreen Lake and southwestern Lake of the Woods (chart numbers 6211, 6240, 6241, 6243, 6247, 6248, 6249, 6251, 6258, 6259, 6260, 6263, 6264, 6267, 6268, 6270, 6271, 6272, 6273, 6274, 6505 and 6506) were therefore digitized, with sounding locations adjusted to positions relative to the shoreline, as defined by NTS 1:250 000 scale topographic maps (Matile and Keller, in prep.). The resulting database contains 31 607 soundings (Fig. GS-29-2). The shoreline was treated as a contour line, and point features depicting rocks and shoals on NTS 1:250 000 scale maps were treated as soundings set to lake level. All data were adjusted for reported chart datum and converted to lake-floor elevation and depth relative to mean lake level. The data have been modelled at a grid resolution of 100 m. The database and model are not intended for navigation.

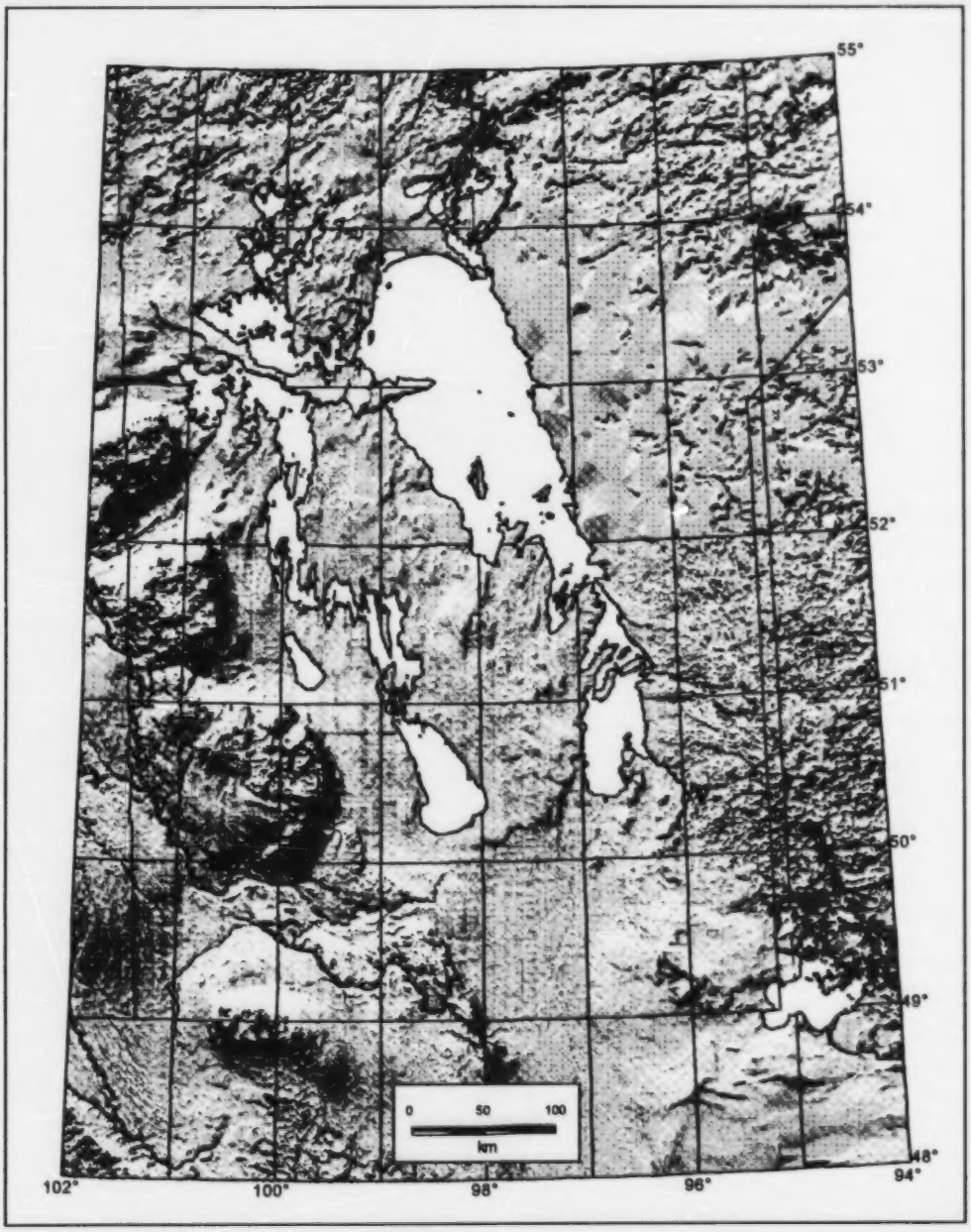

Figure GS-29-1: Shaded-relief depiction of the digital elevation model constructed for the project.

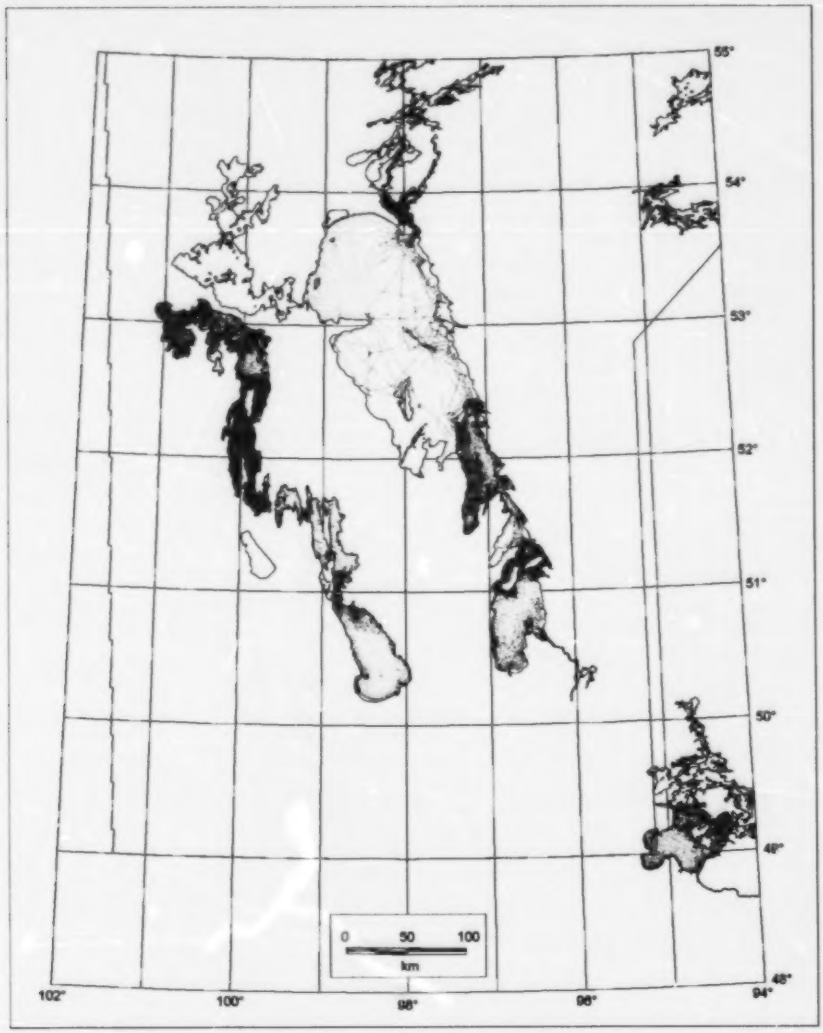

Figure GS-29-2: Location of bathymetric soundings digitized for the project.

Offshore Seismic Surveys

Fine-grained sediments, in places exceeding 100 m in thickness, occur in Lake Winnipeg (Todd et al., 1998, 2000). Similar sediments have been described in Lake Manitoba (Teller and Last, 1981). In order to model this infill in Lake Winnipeg, seismic data collected from the Canadian Coast Guard ship *Namao* in 1994 and 1996 (Todd et al., 1998, 2000; Fig. GS-29-3) are presently undergoing additional processing in order to define surfaces for top of Lake Winnipeg sediments (lake bottom), top of Lake Agassiz sediments, top of till and associated sediments, and top of bedrock. The model is being built with interpolations between seismic lines that are largely guided by bathymetric trends.

Surficial Geology Maps

Surficial geological mapping is used in investigations related to basic science, design and interpretation of geochemical surveys, environmental assessments, engineering applications (e.g. road construction, pipeline integrity, cabling and waste disposal) and management of industrial-mineral resources. It is also urgently required by the agricultural and water-resource sectors as a key layer of information regarding groundwater availability and vulnerability. Surficial geology is therefore an essential layer in the full 3-D model. For the detailed model of the NATMAP study area, recently completed maps at a scale of 1:100 000 (Matile et al., 1998) are being used. Planning is underway for construction of consistent, readily accessible surficial geology for the remainder of southern Manitoba. This next-generation surficial mapping infrastructure will be built, as resources become available, by:

1) digitizing all published maps;
2) building legend equivalency tables in association with various national and international geoscience data-model initiatives;
3) edge-matching seamless 1:250 000 scale map coverage; and
4) generalizing this coverage to a new 1:1 000 000 scale map.

Quaternary Sediments

A model for the Quaternary sediment sequence is being built, largely with reference to the water-well database held by Manitoba Water Resources Branch (GWDrill), guided by surface topography and surficial geology. The model is being constructed using what could be regarded as colour lithological-data scatter plots that depict all drillhole data, in most cases many thousands of sites, within 5 km wide west-east swaths. These scatter plots permit the geologist to apply judgment by occasionally ignoring data based on suspect lithological descriptions or locations apparently in error. The geology is being interpreted using a legend that was built with reference to previous research in the area

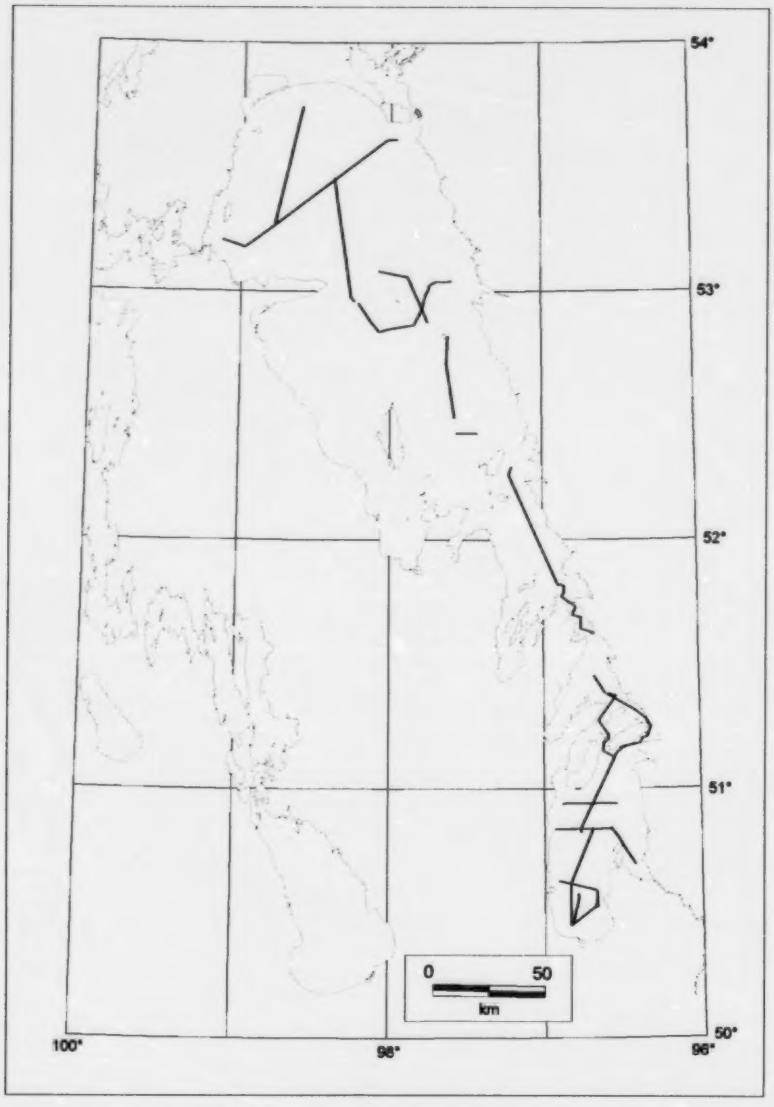

Figure GS-29-3: Location of offshore seismic lines.

(Teller and Fenton, 1980) and rotosonic coring carried out for the purpose of the project (Thorleifson and Matile, 1993). West-east geological cross-sections are being hand drawn, as an overlay on the water-well data. The interpretation is being captured as predicted stratigraphy points at a 5 km spacing, resulting in north-south sections, similar to the west-east fences (Fig. GS-29-4), that act as a check on the interpretation.

Bedrock Surface

A series of bedrock-surface topography maps has been produced by previous workers. These are being updated in association with modelling of the Quaternary sediment sequence.

Bedrock Maps

At present, the 1:1 000 000 scale bedrock geology map of Manitoba (Manitoba Department of Mines, Resources and Environmental Management, 1979) is being used to guide model construction. Plans are in place to increasingly use a complete, seamless Manitoba Geological Survey 1:250 000 scale coverage. Recent efforts, however, have been directed at converting the southern Manitoba Phanerozoic terrane portion of the existing 1:1 000 000 scale bedrock map from its former format to a new format in which polygons are extended over the full extent of the stratum, such that the 2-D model of each stratum can be digitally lifted off, to reveal what is below. This has required reference to previously published Manitoba Geological Survey stratigraphic charts (Bezys and Conley, 1999), as well as other literature that will be described in more detail upon completion of the model.

Phanerozoic Rocks

A complete digital model of the Phanerozoic sequence has been built from previously released Manitoba Geological Survey stratigraphic charts, with some reference to the Western Canada Sedimentary Basin Atlas (Mossop and Shetsen, 1994). The model has more units than were depicted in the database that accompanied the atlas, although fewer than are recognized in the Manitoba Geological Survey Stratigraphic Database, which includes many units only recognizable in limited areas. The current model has been assembled from structure maps for tops of units. Because these surfaces were modelled one at a time, the surfaces,

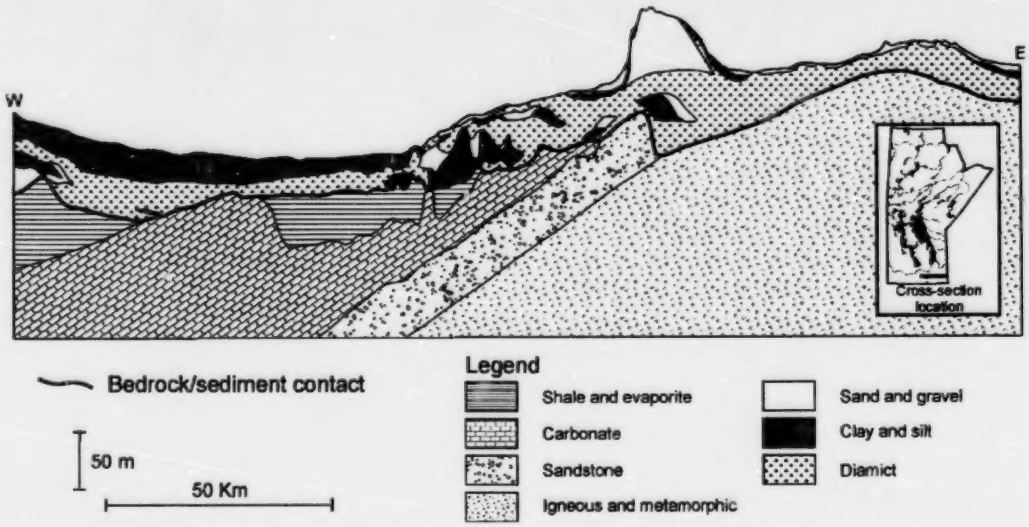

Figure GS-29-4: West-east geological cross-section of Quaternary sediments south of Winnipeg.

as would be expected, wander in data-poor areas. One result of this is that some strata have negative thicknesses in these data-poor areas. Construction of a new model is therefore being contemplated, using as much data as can be prepared in the time available, and with use of multiple surface-modelling procedures.

Sub-Phanerozoic Precambrian Geology Map

A preliminary sub-Phanerozoic Precambrian geology map for all of southern Manitoba Phanerozoic terrane has been prepared for the purpose of this exercise. This information has been compiled from various compilation maps, Project Cormorant and subsurface correlation of drilled Precambrian intervals.

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GS-30 MANITOBA STRATIGRAPHIC DATABASE AND THE CAPITAL REGION STUDY, 2000

by G.G. Conley

Conley, G.G., 2000: Manitoba Stratigraphic Database and the Capital Region Study, 2000; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 194-195.

SUMMARY

In 1999, the northern portion of the Capital Region Study (mineral resource potential, overburden thickness, geology, bedrock topography) was released as preliminary maps (Bezys et. al., 1999a-h). In 2000, the study was extended to the south, and the maps and report for the entire Capital Region Study are presently in the editing stage.

The Manitoba Stratigraphic Database (MSD) continues to be updated with the addition of historic tops to 437 wells.

MANITOBA STRATIGRAPHIC DATABASE AND THE CAPITAL REGION STUDY

In 1999, the Capital Region Study preliminary maps were released. These included the Selkirk (62I/2), Stonewall (62I/3), Teulon (62I/6) and Netley Marsh (62I/7) map sheets (Bezys et al., 1999a-h). In 2000, the study was extended to the south to include the Ste. Anne (62H/10), St. Adolphe, (62H/11), Winnipeg (62H/14) and Dugald (62H/15) map sheets. Included are overburden thickness, geology, bedrock topography (5 m contour interval) and mineral potential maps (*see* Bezys, GS-31, this volume). Data sources include 5897 water wells and 182 stratigraphic wells from the MSD. Production of the maps and report is in progress and release is expected in March 2001.

When the 1999 preliminary maps were prepared, the best reference for ground elevation was the existing 1:50 000 NTS map sheets, with a contour interval of 25 feet. Because of the coarseness of the maps, assigning ground elevations was difficult. As a result, when a digital elevation model (DEM) was produced, the resultant surface contained many anomalous peaks and pits.

Late in 1999, Manitoba Industry, Trade and Mines acquired a set of 1:20 000 digital map tiles, centred on the City of Winnipeg. The maps contained both 2.5 m contours and numerous spot elevations. With the availability of this new and very detailed ground elevation data, it was decided to check and correct the ground elevations for all water-well data gathered prior to the completion of the final maps.

To accomplish the correction, all of the individual 1:20 000 map tiles were combined in ArcView® to form a topographic map covering the study area. Likewise, the spot elevations were combined to form a continuous data set. The Capital Region well data were then compared to the 1:20 000 contour and spot data. Wells with ground elevation deviations of greater than 0.3 m were corrected. To check progress at various stages, DEMs were prepared of the Capital Region data and then compared to the 1:20 000 contours.

Once the ground elevation was satisfactorily corrected, the depth to bedrock data also needed to be evaluated. Because of the large number of wells used in this study, the wells tend to be tightly grouped throughout most of the study area. When the depth to bedrock of a single well deviated significantly from those of its neighbours, it was usually assumed to be incorrect and removed from the data set. When two or more adjacent wells exhibited similar anomalous depths, the water-well stratigraphy was called up in ArcView® and examined. If the stratigraphy appeared to be correct, the data points were kept; otherwise, they were removed.

The original water-well data gathered for the Capital Region Study were selected because they had reasonably accurate descriptive locations. The wells were plotted from the descriptions and UTM co-ordinates generated by digitizing or by measurement. However, some parts of the study area had large gaps with little or no data. As the gaps cause problems in the generation of contour or DEM maps, it was necessary to add infill data. This introduced another problem, as the co-ordinates in the Water Resources water-well database only located the well to the centre of a legal subdivision. Even in lightly populated areas, this frequently results in two or more points being mapped to the same co-ordinate. In more highly populated areas, many wells would display at a single co-ordinate, with depths varying by many tens of metres. The procedure used to select the infill well was to examine the stratigraphy of each water well at the location and then select the most appropriate one based on the stratigraphic characteristics of the nearest neighbours. Stratigraphic wells from the Manitoba Stratigraphic Database were used as key reference points.

Once the data-editing process was complete, the well data were inserted into AutoCAD® 14 and contoured using QuickSurf®. The maps and report are now in the editing stage and are expected to be released in March 2001.

MANITOBA STRATIGRAPHIC DATABASE – UPDATE

The Manitoba Stratigraphic Database continues to be updated. This has proceeded more slowly than expected, however, due to the amount of time required by the Capital Region Study. At this time, 437 stratigraphic and/or oil and gas wells have been updated by historical tops picked by H.R. McCabe (formerly of Manitoba Industry, Trade and Mines). An update of the MSD CD-ROM is planned for middle to late 2001. More detailed information on both the Manitoba Stratigraphic Map Series and Manitoba Stratigraphic Database is available in Conley and Bezys (1998) and Bezys and Conley (1999).

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SUMMARY

Stratigraphic investigations and drilling programs were carried out for various projects. The Capital Region Study has been completed and the report is in the editing stage. H.R. McCabe's historical corehole data have been added to the Manitoba Stratigraphic Database. All brine springs are being tabulated for the Prairie-Type Project. Two coreholes were drilled in the Dancing Point (Lake Winnipeg) area for stratigraphic purposes; one corehole was drilled proximal to the High Rock Lake structure; one corehole was drilled in the Wekusko area to test metallic veining in the Red River Formation; and one corehole was drilled in the Winnipegosis area for regional stratigraphy. A total of 457.8 m of drilling was conducted this year.

STRATIGRAPHIC INVESTIGATIONS

Capital Region Project

A mineral-resource and land-use assessment of Manitoba's Capital Region (Winnipeg and portions of the surrounding municipalities) was undertaken by the Manitoba Mines Branch and the Manitoba Geological Survey. The study is being conducted in response to the Capital Region Strategy, under development by the Manitoba Round Table. The purpose of this assessment is to provide mineral-resource data for use in municipal development plans for the Capital Region that will legally protect high-quality quarry minerals, such as crushed stone (Bamburak and Bezys, 1995, 1996).

In 1999, eight preliminary maps were released that depict overburden thickness, bedrock topography and mineral resource potential in NTS map sheets 62I/2, /3, /6 and /7 (Bezys, Bamburak and Conley, 1999a–h). Final versions of these maps, as well as eight maps covering NTS map sheets 62H/10, /11, /14 and /15, are in the final stages of preparation; the accompanying economic report will be released for March 2001 (*see* Conley, GS-30, this report).

Manitoba Stratigraphic Database

The Manitoba Stratigraphic Database continues to be updated. At this time, 437 stratigraphic and/or oil and gas wells have been updated with historical tops picked by H.R. McCabe (*see* Conley (GS-30, this volume) and Bezys and Conley (1999) for further information).

Prairie-Type Microdisseminated Mineralization

This project is the investigation of Prairie-type microdisseminated mineralization in Devonian-age carbonate rocks in the Mafeking quarries (west-central Manitoba; Fedikow et al., 1996). Surface mapping is augmented by geochemical and geophysical surveys: brine spring, b-horizon soil and stream sampling. Rock outcrops in the Mafeking quarries display unusual 'chimney' features that appear to represent hydrothermal vents. Palynological (conodont) analyses of sand and clay infill from the centre of these chimneys indicate a Devonian Souris River Formation age, the same as that of the carbonate host rock of the chimneys. There is no evidence of Cretaceous-age infill to suggest a karst origin. Sideritic rinds adjacent to the chimneys are slightly enriched in some elements, such as Co, Ni, Zn and Fe, when compared with surrounding host rock strata (Table GS-31-1). This enrichment suggests that mineralizing fluids were coming from underlying strata as brines or formation waters.

An open file report is in preparation and is expected to be released in March 2001. A B.Sc. thesis on the Mafeking quarry has been completed in conjunction with the University of Manitoba (Ramnath, 1999). Nuno Machado (University of Montreal) is carrying out age-dating work on the zircon and monazite. Steve Grasby (Geological Survey of Canada (GSC–Calgary) continues to conduct sample analyses

Table GS-31-1: Select INAA geochemistry of solution chimneys from the North Mafeking quarry, west-central Manitoba.

Element	Co	Fe	Ni	Zn
Units	ppm	%	ppm	ppm
Detection Limit	1	0.01	20	50
SC1:				
Sinter	-1	0.19	-20	-50
Rind	140	36.8	88	577
Wall rock	2	0.48	-20	-50
SC2:				
Sinter	1	0.37	-20	-50
Rind	17	32.2	-20	188
Wall rock	2	1.31	-20	-50
SC3:				
Sinter	1	0.28	-20	-50
Rind	19	32.9	-20	174
Wall rock	2	0.85	-20	-50
SC4:				
Sinter	1	0.21	-20	65
Rind	270	38.4	-20	296
SC7A:				
Sinter	2	0.26	-20	-50
Rind	43	27.2	-20	480
SC7B:				
Sinter	1	0.19	-20	-50
Rind	49	26.5	-20	588
Wall rock	2	0.61	-20	-50
SC8:				
Sinter	2	0.18	-20	-50
Rind	10	27.7	-20	67
Wall rock	2	1.14	-20	-50
SC9:				
Sinter	1	0.19	-20	-50
Rind	110	28.1	71	640
SC10:				
Sinter	2	0.26	-20	-50
Rind	87	25.5	60	472
SC11:				
Sinter	12	0.68	-20	-50
Sinter	7	0.57	-20	-50
Rind	190	31.1	89	193
Wall rock	4	1.41	-20	-50
SC12:				
Sinter	1	0.18	-20	-50
Rind	72	40.7	-21	158
SC13:				
Sinter	2	0.13	-20	-50
Rind	8	28.6	-20	99
SC14:				
Sinter	1	0.14	-20	-50
Wall rock	3	2.71	-20	-50
SC15:				
Sinter	2	0.2	-20	-50
Rind	33	29.6	75	184
SC17:				
Sinter	2	0.3	-20	-50
Rind	37	34.5	-20	80
Wall rock	1	0.71	-20	-50

of brine springs for stable-isotope geochemistry (*see* Grasby, GS-34, this volume). All brine springs in west-central Manitoba are being inventoried and will be presented as a GSC bulletin.

STRATIGRAPHIC AND INDUSTRIAL MINERALS COREHOLE DRILLING

Dancing Point

Two coreholes were drilled in the Dancing Point area, east of Highway 6 and west of the Lake Winnipeg shoreline (Fig. GS-31-1; Table GS-31-2). Corehole M-1-00 went to 170.5 m and intersected the Precambrian; corehole M-2-00 went to a depth of 151.8 m and intersected the Winnipeg Formation sandstone. These coreholes were drilled in an area where silicified coral heads (5–10 cm in diameter) were found along the Silurian escarpment (W.D. McRitchie, pers. comm., 2000). Corehole M-1-00 confirmed the stratigraphic location of these coral heads, from within the middle of the Stonewall Formation. Corehole M-2-00 did not encounter any evidence of the silicified coral heads. These two coreholes will be valuable reference holes for NTS map sheet 63B, where very few coreholes have been drilled.

High Rock Lake

One corehole, M-3-00, was drilled east of the High Rock Lake structure (Fig. GS-31-2, -3; Table GS-31-2). A study conducted by H.R. McCabe indicated the presence of a highly disturbed area, possibly

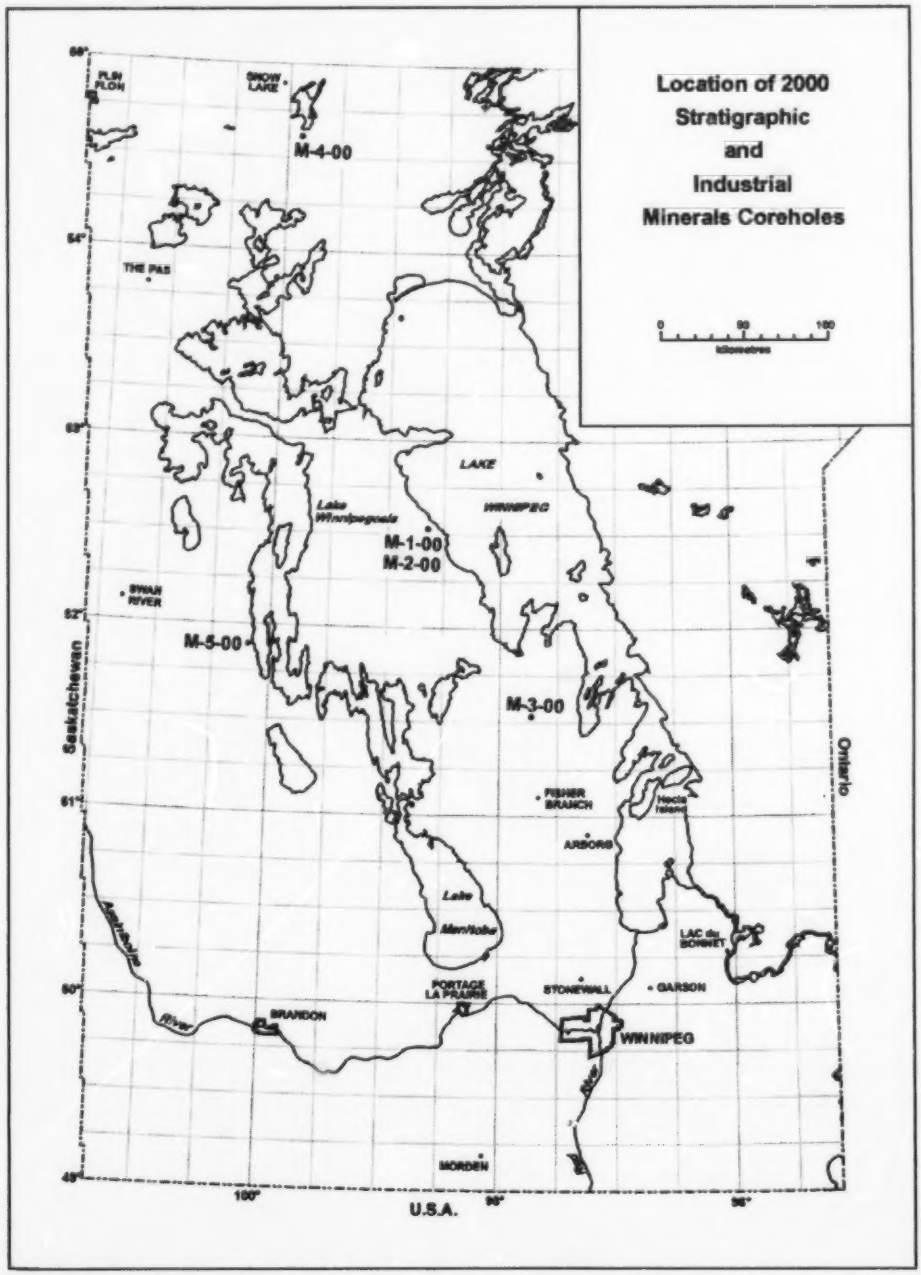

Figure GS-31-1: Location of stratigraphic and industrial-mineral coreholes, 2000.

Table GS-31-2: Summary of stratigraphic corehole data, 2000.

Hole no.	Location and elevation (m)	SYSTEM/Formation/ (Member)	Interval (m)	Lithology summary
M-01-00	02-03-41-09W	OVERBURDEN	0.0–3.0	Glacial till
Dancing Pt.	5815850N	SILURIAN/Interlake Group/	3.0–7.8	Buff wackestone; very broken core; no *Virgiana*
	522125E	Fisher Branch		
	243.8	ORDOVICIAN-SILURIAN/	7.8–33.5	Brown mudstone to wackestone, some chert
		Stonewall		7.8–8.0: Stonewall Marker; light brown mudst.
				10.9–13.0: T-Zone; dark grey to some yellow, grainstone to mudstone
		ORDOVICIAN/Stonewall		13.0–27.6: light brown, mottled, fractured with silicified corals
		(Williams)		27.6–33.5: dark grey mudstone, laminated
		Stony Mountain	33.5–67.5	Light brown wackestone, mottled
		Red River (Fort Garry)	67.5–88.0	Grey and brown mudstones, laminated,some chert
		(lower Red River)	88.0–150.2	Brown wackestone, mottled, cherty
		Winnipeg	150.2–168.4	150.2–156.6: dark grey to grey sandstone, some shale, burrowed
				156.6–159.2: Light grey, fine grained sandstone
				159.2–162.3: 10 cm of fine-grained brown sandstone (?infill)
				162.3–165.3: light grey, fine-grained sandstone (10 cm of core)
				165.4–168.4: no core
		PRECAMBRIAN	168.4–170.5	168.4–169.4: Regolith: white to grey, fine- to coarse-grained, severely kaolinized, unconsolidated rock
				169.4–170.5: Hornblende granodiorite: light grey to green, coarse- to very coarse grained, poorly foliated to massive rock; composition: hornblende 10–30% (1–10 mm in size), quartz and feldspar
M-02-00	15-06-41-09W	SILURIAN/Interlake Group/	0.0–4.2	Buff wackestone and packstone, abundant *Virgiana*
Dancing Pt.	5816175N	Fisher Branch		
	517125E	ORDOVICIAN-SILURIAN/	4.2–30.2	4.2–5.4: Stonewall Marker: brown mudstone, broken core
	237.7 m	Stonewall		5.4–6.8: buff mudstone
				6.8–10.4: T-Zone: blue-grey mudstone, burrowed
				10.4–21.0: buff wackestone, porous
				21.0–21.5: blue-grey marker bed, mudstone
				21.5–24.5: buff wackestone and mudstone
		(Williams)		24.5–30.2: brown to grey, laminated mudstone
		Stony Mountain	30.2–63.9	Buff mottled wackestone
		Red River/(Fort Garry)	63.9–82.5	Interbedded laminated mudstone, brown to grey
		(lower Red River)	82.5–146.5	Mottled wackestone, burrowed
		Winnipeg	146.5–151.8	146.5–149.5: Green-grey to black-grey shale with some siltstone, burrowed
				149.5–151.8: White quartzose sandstone, friable at base, becoming consolidated at top
M-03-00	16-20-28-01W	ORDOVICIAN/	0.0–11.9	Blue-grey laminated mudstone
High Rock Lake	5697600N	Red River (lower)		
	606250E	(Fort Garry)	11.9–98.7	Buff mottled wackestone (Tyndall Stone), cherty, very porous
	225 m			
		Winnipeg	98.7–127.2	98.7–103.8: green mudstone, some sand, sharp contact
				103.8–114.0: brown to grey mudstone (shale) and siltstone, burrowed
				114.0–127.2: white quartzose sandstone
M-04-00	05-20-65-16W	ORDOVICIAN/	0.0–0.85	Beige mottled dolomite
Wekusko Lake S	6054470N	Red River (Hecla Beds)		

Table GS-31-2: Summary of stratigraphic corehole data, 2000. ***(continued)***

Hole no.	Location and elevation (m)	SYSTEM/Formation/ (Member)	Interval (m)	Lithology summary
	445016E	Winnipeg	0.85–0.92	Sandstone, pyritic with dolomite clasts
	266.7 m	PRECAMBRIAN	0.92–14.35	0.92–5.25: light green-grey sericite schist, vertically sheared (missing 2 m of core)
				5.25–7.35: Granite schist, steeply dipping
				7.35–8.45: Soft sericite schist (only 0.27 m of core)
				8.45–14.35: Steeply dipping granite schist
M-05-00	01-05-33-19W	DEVONIAN/Souris River	0.0–20.2	Fine-grained dolomite
Pine River	5739324N	(Sagamace)	20.2–27.1	Dolomitic shale, limestone and breccia
	423445E	(Point Wilkins)	27.1–45.9	Dolomite, brecciated
	262 m	(First Red Beds)	45.9–56.6	Shale breccia with disturbed bedding
		Dawson Bay		
			56.6–69.8	Limestone, some dolomite
			69.8–82.8	Calcareous shale, green-grey to purple
			82.8–94.8	Limestone, brachiopod rich
		(Second Red Beds)	94.8–106.7	Polymict collapse breccia
		Winnipegosis	106.7–112.7	Dolomite breccia, fossiliferous
		(Transition Beds)		
		(Upper)	112.7–117.1	Dolomite, bituminous, black mineralization (sphalerite)
		(Lower)	117.1–121.1	Platform Facies, dolomite, vuggy and nodular

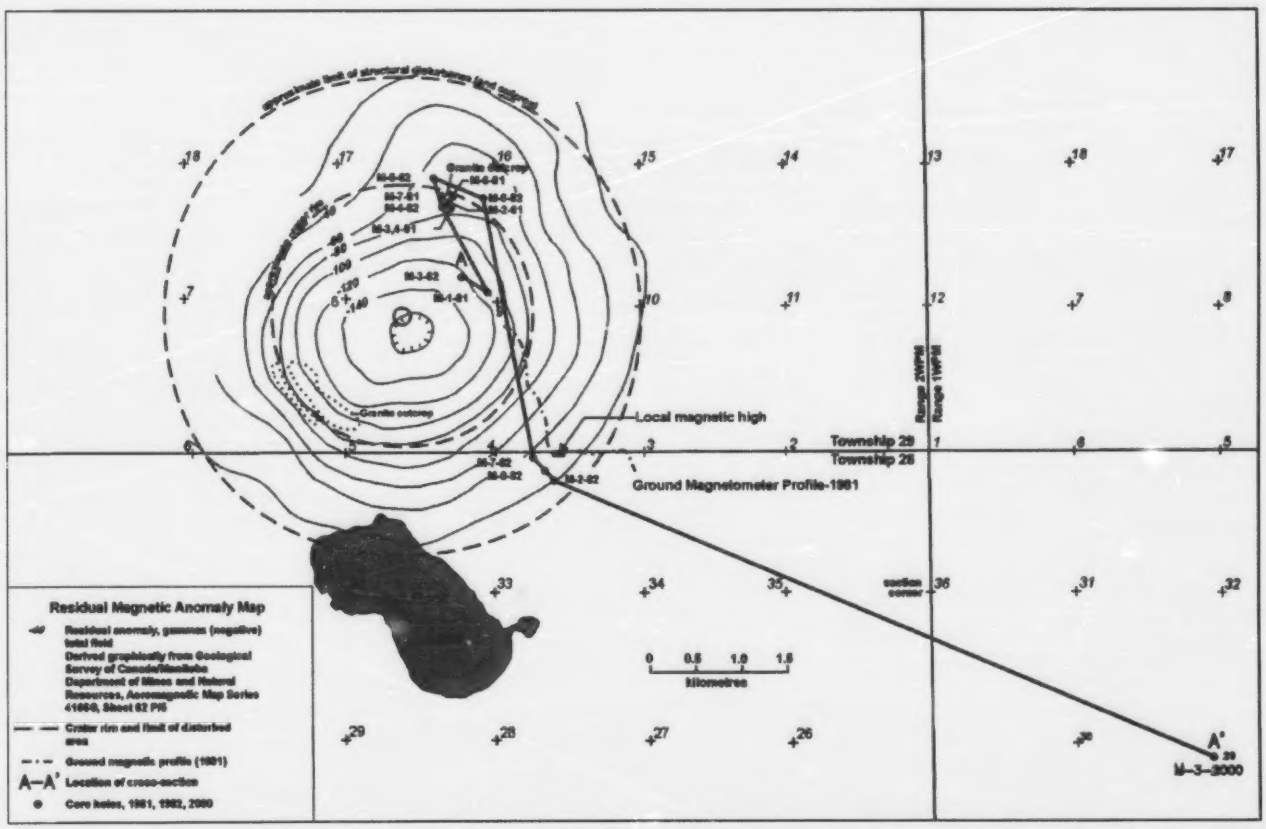

Figure GS-31-2: Location of coreholes in the High Rock Lake structure magnetic map and of cross-section A-A' (after McCabe, 1982).

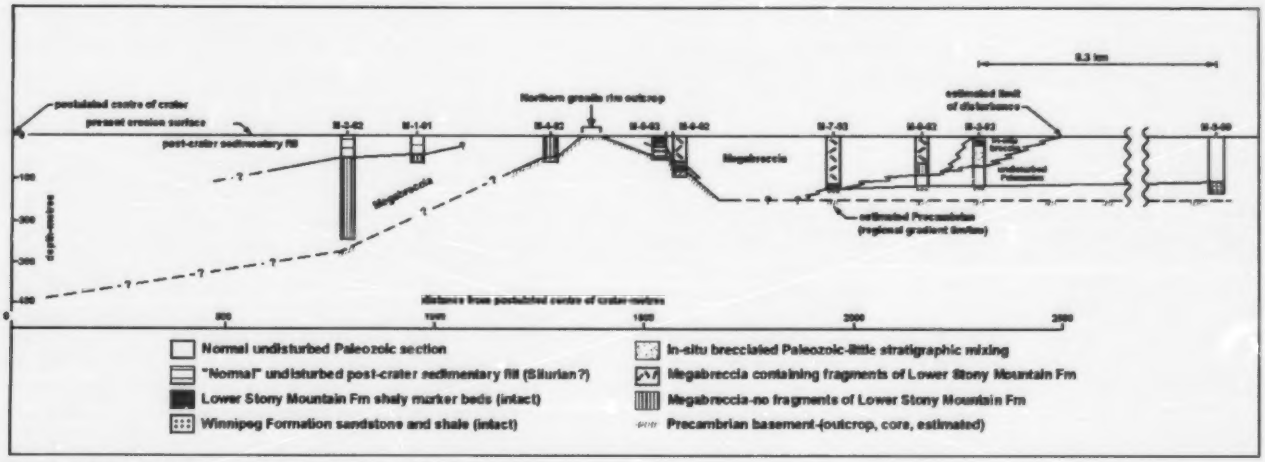
Normal undisturbed Paleozoic section
"Normal" undisturbed post-crater sedimentary fill (Silurian?)
Lower Stony Mountain Fm shaly marker beds (intact)
Winnipeg Formation sandstone and shale (intact)
In-situ brecciated Paleozoic-little stratigraphic mixing
Megabreccia containing fragments of Lower Stony Mountain Fm
Megabreccia-no fragments of Lower Stony Mountain Fm
Precambrian basement-(outcrop, core, estimated)

Figure GS-31-3: Cross-section A-A' (after McCabe, 1982).

representing a crypto-explosion meteorite-impact structure (McCabe, 1981). Two Precambrian granite inliers are approximately 180 m above their expected position in the regional structure. As well, disturbed Paleozoic outcrops exist north and west of the granite outcrops. The structurally disturbed area is shown to be coincident with a well defined aeromagnetic low (McCabe, 1981).

Corehole M-3-00 was drilled to provide an undisturbed reference hole in the High Rock Lake area. It was hoped that the hole would reach the Precambrian, but it encountered unconsolidated sand in the Winnipeg Formation and drilling had to be terminated prematurely. The hole is undisturbed and does not appear to be affected by the High Rock Lake structure (Fig. GS-31-3).

Wekusko Lake South

In July 1999, mineralized fractures were noted within Ordovician Red River Formation dolomite in a Manitoba Highways quarry (UTM 14, E445016, N6054470), located south of Wekusko Lake and Provincial Trunk Highway 39. Located near the east wall of the quarry, two parallel fractures, trending 085°, are spaced 3 m apart and are intersected by another fracture trending 170°. Mineralization occurs in the dolomite along the trend of the fractures over a width of 10 cm on either side of the fractures. The mineralization appears as vuggy, mauve to reddish brown or dark grey clots and sinuous stringers, with minor bright blue-green– (bornite and azurite) and brass-coloured (pyrite) crystals that appear along subfractures and within healed dendritic tendrils. A sample was sent to Activation Laboratories Ltd. for analysis by neutron activation and ICP in 1999. The values indicate a relative enrichment in Pb, Hg, Cu, Sr, As and Co. An additional sample was sent to the same laboratory early in 2000, and the results confirmed the enrichment.

Corehole M-4-00 was drilled to a depth of 14.4 m to intersect the mineralization, but only thin clots and veinlets were found near the top of the hole. A total of 13.4 m of Precambrian was intersected, a highly foliated sericite schist.

Camperville Gravity Low

The Pine River Junction corehole M-5-00 was drilled through 121.0 m of Devonian stratigraphy to verify the presence of sphalerite originally encountered in corehole M-6-80. The honey-coloured sphalerite was found in the Upper Member of the Winnipegosis Formation (McCabe, 1980; Gale and Conley, 2000). In hole M-5-00, black mineralization was encountered at the same stratigraphic level as in hole M-6-80, and appears to be the 'blackjack' form of sphalerite. No honey-coloured sphalerite was found in hole M-5-00. For further information, see Bamburak et al. (GS-32, this volume).

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GS-32 INVESTIGATIONS ON THE CAMPERVILLE GRAVITY LOW (PARTS OF NTS 62N/E AND 63C/E)

by J.D. Bamburak, I.T. Hosain, M.A.F. Fedikow and E. Dobrzanski[1]

Bamburak, J.D., Hosain, I. T., Fedikow, M.A.F. and Dobrzanski, E. 2000: Investigations on the Camperville gravity low (parts of NTS 62N/E and 63C/E); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 202-209.

SUMMARY

Over many years of documentation by the Manitoba Geological Survey and the Geological Survey of Canada, anomalous features have been found in the vicinity of Camperville, Manitoba. Other recent studies have concluded that the Superior Boundary Zone, lying immediately to the northwest, has been active during the Paleozoic. A major fault appears to separate the Superior Boundary Zone from the Camperville gravity low. Long-term movements along this fault may have produced channelways for deep fluid migration from the buried Precambrian basement. These channelways may have also penetrated the overlying Phanerozoic strata. The fluids may have left their signature by depositing minerals at structural and lithological traps. The purpose of this summer's field work was to determine if unique mineralization is present in the vicinity of the Camperville gravity low, and if this mineralization has economic potential.

INTRODUCTION

The Camperville area contains numerous anomalous features that have been documented by the Manitoba Geological Survey, Geological Survey of Canada and other agencies. The most conspicuous of these is a semicircular (65 km diameter) gravity low of approximately 30 mGal, the most intense low in the province (Fig. GS-32-1). This gravity low was noted in the mid-1960s (Buck, 1968). The Camperville gravity low (CGL) is abruptly truncated on the northwest by the northeast-trending gravity high of the Superior Boundary Zone (SBZ), which roughly parallels the Swan River.

[1] Manitoba Museum of Man and Nature, 190 Rupert Avenue, Winnipeg, Manitoba R3B 0N2

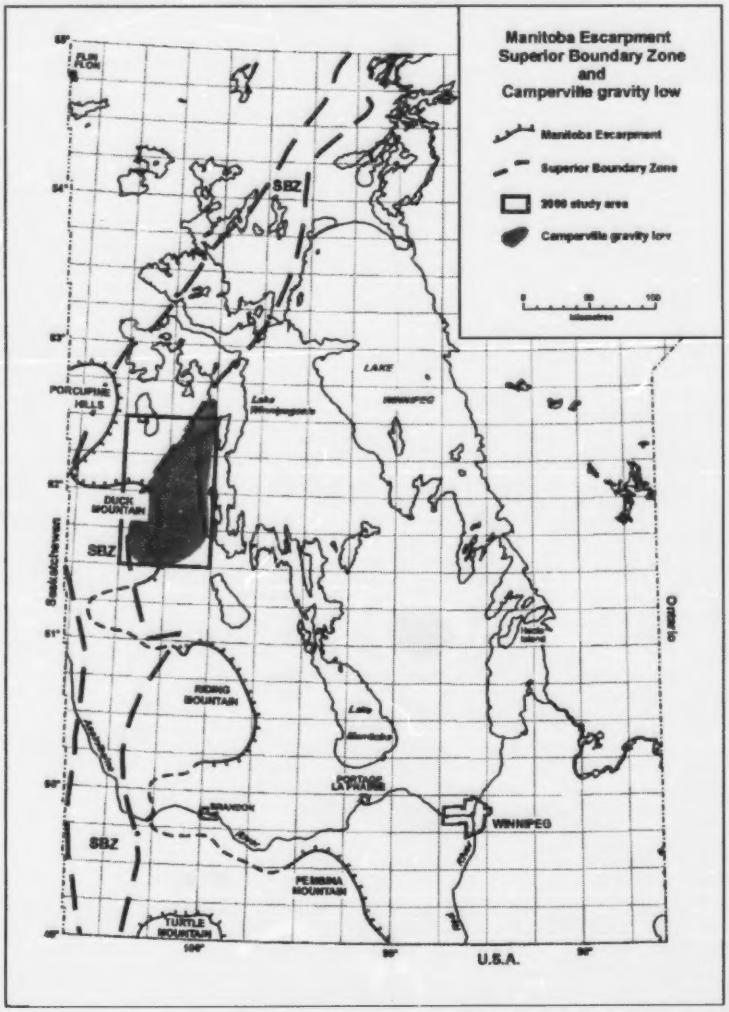
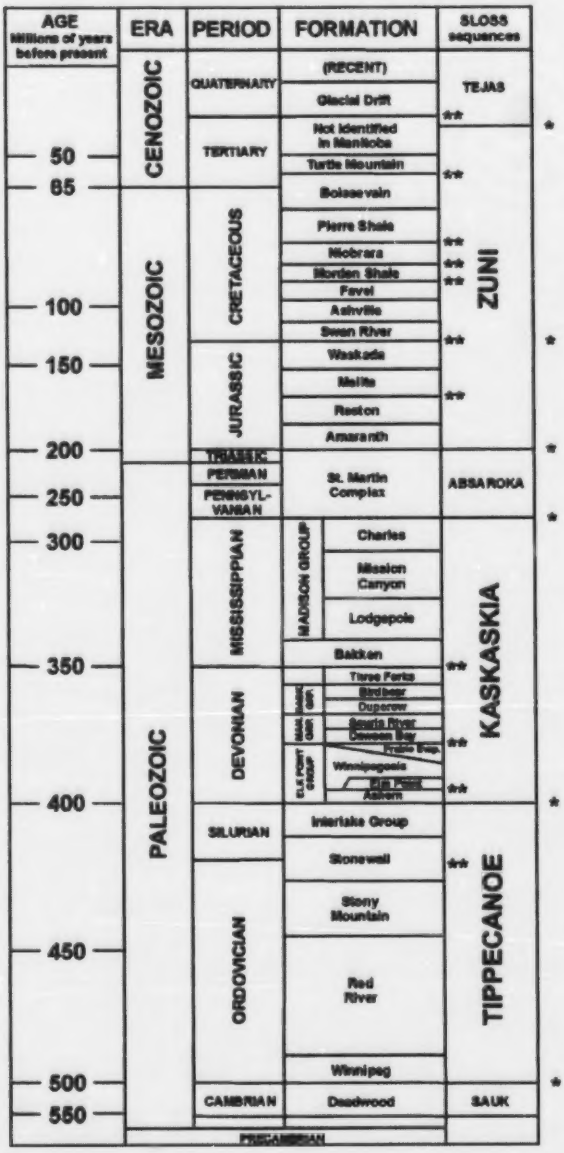


Figure GS-32-1: Location of the Camperville gravity low (CGL) study area, the Manitoba Escarpment and the Superior Boundary Zone (SBZ) in southern Manitoba.

Stratigraphy

The stratigraphic succession of deposition and erosion of formations in southern Manitoba is shown in Figure GS-32-2. The Devonian and Upper Cretaceous stratigraphy of the CGL study area is shown in more detail in Table GS-32-1. For the most part, the Upper Cretaceous marine shale and thin limestone beds were removed by erosion east of the Manitoba Escarpment (Fig. GS-32-1).



Footnote: * Potential major karst events
** Potential minor karst events

Figure GS-32-2: Generalized stratigraphic section for southern Manitoba, with asterisks showing times of karst events (McRitchie, 1991).

The oldest exposed bedrock within the CGL is the Devonian Winnipegosis Formation, located east of the Manitoba Escarpment (Fig. GS-32-1). The Winnipegosis Formation is overlain by the Dawson Bay Formation, which is overlain by the Souris River Formation of the Manitoba Group. Unconformably overlying the Devonian beds is the Upper Cretaceous Swan River Formation. The basal formation of the Manitoba Escarpment is the Ashville Formation, followed in ascending order by the Morden Shale, Niobrara Formation, and the Gammon Ferruginous, Pembina, Millwood and Odanah members of the Pierre Shale (Table GS-32-1).

Pleistocene glacial sediments blanket the Camperville gravity low.

Anomalous Features

Other anomalous features within the Camperville gravity low and/or in its immediate vicinity are:

1) Numerous circular aeromagnetic highs, contained within the -60 mGal contour of the CGL.
2) A residual magnetic high of more than 1000 gammas within the northern part of the CGL, 10 km north of Provincial Trunk Highway (PTH) 20 and northwest of Camperville (Hall et al., 1975).
3) An area of 'magnetic disturbance', situated 2 km northeast of Snake in the south basin of Lake Winnipegosis, 10 km east-northeast of the town of Winnipegosis (Radakir, 1973).
4) A change in orientation of the SBZ from northeast-southwest to north-northwest–south-southeast.
5) The boundary between the sand and shale facies within the Ordovician Winnipeg Formation.
6) The apparent thinning of the Ordovician Red River Formation isopach and a corresponding apparent thickening of the overlying Silurian Interlake/Stonewall isopach.
7) The northern edge of the known Jurassic beds in southern Manitoba.
8) The eastern erosion edge of Mesozoic shale beds of the Manitoba Escarpment.
9) The presence of honey-coloured sphalerite, with minor pyrite, identified in Winnipegosis Transition Zone core between 110.8 and 113.2 m in hole M-6-80, drilled in the vicinity of a quarry on Pine River Road, SE1-5-33-19W (McCabe, 1980).
10) Barite, identified by McCabe (1980), in lower Dawson Bay beds within a quarry in NW16-31-33-19W (10 km north of hole M-6-80).
11) A structurally high, flat-topped, Winnipegosis reef at a depth of 21 m in corehole M-9-79, drilled in the vicinity of the above quarry. With an estimated thickness of 105 m, the reef is the thickest reef known in this portion of the outcrop belt. A strong artesian flow (20 m head) was also encountered (McCabe, 1979, 1980). A value of 200 ppm Pb was obtained from reefoid dolomite (with excellent porosity) of the Winnipegosis Formation at a depth of 30.5 m (Gale and Conley, 2000).
12) A dispersion train of five G10 garnets, trending southeast from the CGL for a distance of 400 km. The nearest G10 garnet to the anomaly was found only 70 km to the southeast (Thorliefson and Garrett, 1993). A sixth G10 garnet was found within the train in a subsequent survey (Matile et al., 1996).

Previous Investigations

The potential for Mississippi Valley–type (MVT) mineral occurrences within Paleozoic carbonate rocks in Manitoba has been suggested by many geologists over the past 30 years (Gale and Conley, 2000). Fedikow (1999) discussed the potential for sediment-hosted disseminated sulphide deposits (Au, platinum-group elements and base metals).

An oil-well testhole was drilled in the area of Paradise Beach, southeast of Winnipegosis (Fig. GS-32-3) in the 1930s or 1940s. The well was on a dome-shaped outcrop of Dawson Bay Formation. The remains of the drill camp can still be seen near the shore of Lake Winnipegosis. Local stories of the drilling project tell of an underground exploration shaft for lignite, base metals or rare elements.

In 1969, a small, well rounded pebble of almost pure galena was found in NE26-14-1E, near the town of Balmoral, 250 km southeast of the CGL and 35 km northeast of Winnipeg. McCabe (1969) suggested that the pebble might have been glacially transported from a site similar to the Phanerozoic Pine Point deposit (NWT).

Table GS-32-1: Table of formations in the vicinity of the Camperville gravity low.

Formation/Member	Maximum thickness (m)	Lithology
Upper Cretaceous		
Pierre Shale		
Odanah Member	150	Hard grey siliceous shale
Millwood Member	60	Soft bentonitic clay
Pembina Member	7	Noncalcareous black shale with numerous bentonite interbeds near base
Gammon Ferruginous Member	30	Ferruginous black shale
Niobrara Formation	30	Chalky buff and grey speckled calcareous shale
Morden Shale	30	Noncalcareous black shale with abundant jarosite
Favel Formation		
Assiniboine Member	17	Olive-black speckled calcareous shale with Marco Calcarenite beds near top
Keld Member	17	Olive-black shale, speckled shale, with Laurier limestone beds near top
Ashville Formation	80	Noncalcareous black to dark grey shale, silty; Newcastle sand zone, in places
Swan River Formation	150	Sandstone, sand and silt, quartzose, pyritic shale, noncalcareous
Devonian		
Souris River Formation		
Sagamace Member	34	Basal, red and orange mottled, calcareous and noncalcareous shale; argillaceous limestone; brecciated dolomite and limestone; aphanitic, partly vuggy limestone with abundant sparry calcite
Point Wilkins Member	30	Basal red shale; argillaceous and high calcium limestone
Dawson Bay Formation	50	Basal red shale; dolomite; lower limestone, fossiliferous; medial calcareous shale; dolomite; upper stromatoporoidal limestone
Winnipegosis Formation	28 (105)[a]	Dolomite, reefoid; limestone, interreef bituminous laminates
Ashern Formation	26	Dolomite, argillaceous dolomite; dolomitic shale

[a]See item 11 in 'Anomalous Features' section of text.

In 1975, Gulf Minerals Canada Ltd. conducted a systematic search for base-metal deposits along the rim of the Western Canada Sedimentary Basin. Ten claim blocks (6752-6761) were staked in August 1975, near the town of Minitonas (Fig. GS-32-3). Gulf Minerals selected the Minitonas area as an exploration target based on:

1) the projected trend (from aeromagnetic data) of the SBZ beneath Paleozoic and Mesozoic cover;
2) the distribution of Winnipegosis reefs;
3) post-Devonian solution of the Prairie Evaporite and probable collapse structures;
4) the Silurian Interlake Group unconformity with the overlying Middle Devonian Ashern Formation and the possible development of karst topography; and
5) the presence of circular structures on the Earth Resources Technology Satellite (ERTS) mosaic.

During November 1975, Gulf Minerals conducted a four-hole drilling program on the claims in accord with Agreement 24 with the Province of Manitoba. The holes drilled in twp. 36, rge. 25 and 26, W1 ranged from 508.4 to 535.5 m in depth, for a total of 2087.6 m. Of this total, 1481.3 m of the core were analyzed, over 3 m lengths, by

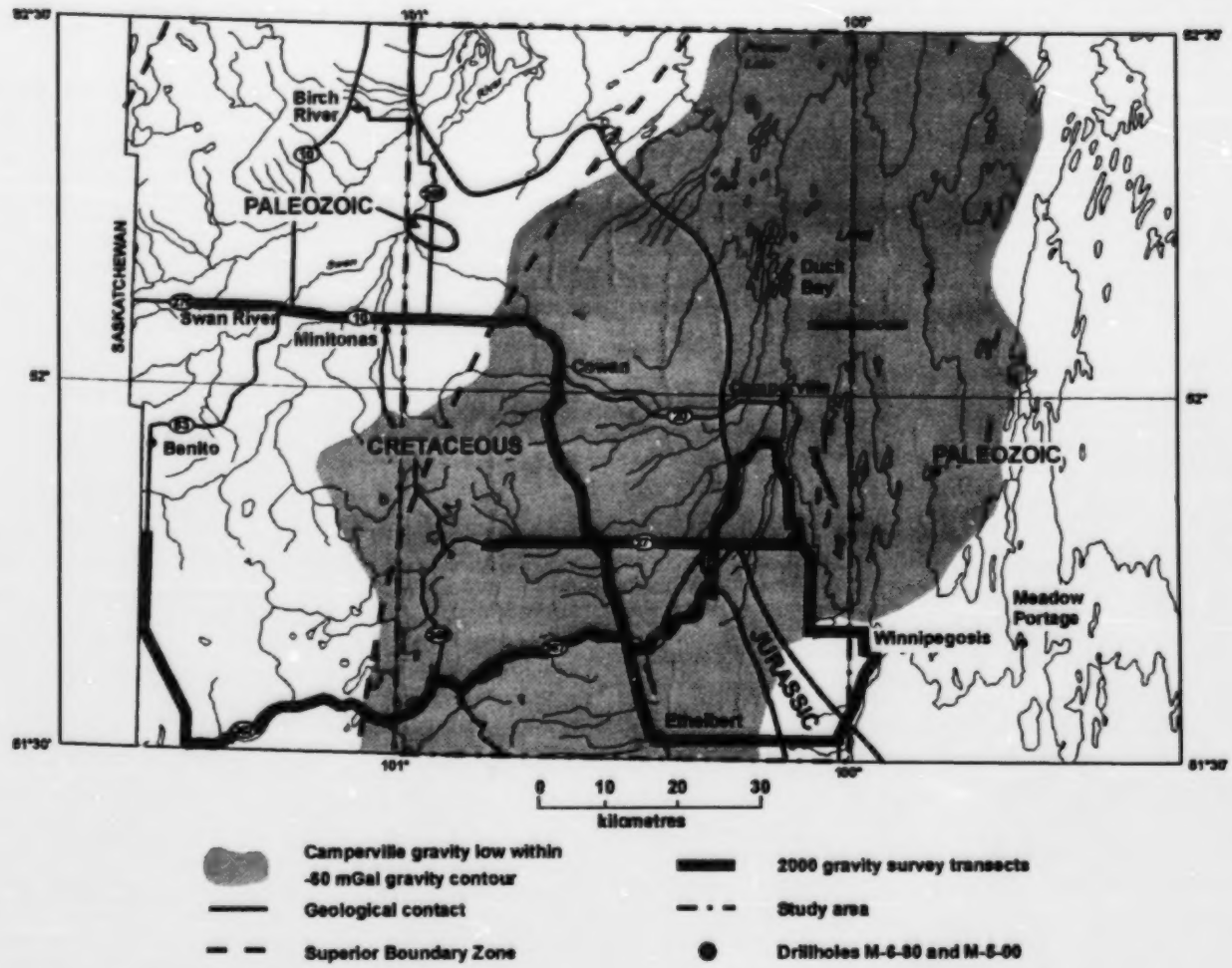


Figure GS-32-3: Camperville gravity low study area.

Technical Services Laboratories, Toronto. The samples were analyzed for Cu, Pb and Zn by the atomic absorption method (Assessment File 92116, Manitoba Industry, Trade and Mines, Winnipeg).

Pyrite was found as fillings in fractures and vugs, as well as being associated with fossils. A small crystal of sphalerite was noted at 429.2 m (1,408 feet) in hole 13-10-36-26W1 (Assessment File 92116). The maximum values for Cu, Pb and Zn occur within a sandy and shaly bed in the Interlake Group. The bed is present in three of the drillholes, and the metal content increases toward the north. The highest Cu value from the Winnipegosis Formation (182 ppm) is from the thickest section in hole 1-28-36-26W1. Another interesting interval is from hole 15-32-36-26W1, where an average of 297 ppm Cu and 294 ppm Zn were obtained from 341.4 to 353.6 m (1,120 to 1,160 feet) in the Interlake Group, with a maximum of 721 ppm Cu and 690 ppm Zn from 347.5 to 350.5 m (1,140 to 1,150 feet). However, Gulf Minerals concluded that there is no significant base-metal content in the sampled intervals and that further work was not warranted (Assessment File 92116).

The results of a detailed geochemical sampling program, carried out from 1980 to 1985 on departmental drill core, was released by Gale and Conley (2000). The utilization of data from this report is described below (*see* 'Geological Investigations' section).

Recent Investigations

Oil-well core and chips, and outcrop samples, from the southern portion of the Manitoba Escarpment were analyzed by Fedikow et al. (1997) to provide new metallogenic concepts relevant to mineral exploration in Phanerozoic sequences. The results of these analyses were published as a series of data files (Fedikow et al., 1998). Additional black-shale samples were collected in the Porcupine Hills area as part of the Prairie-type microdisseminated mineralization study centred on the Mafeking area (Bamburak et al., 1997). In 1999, this work was extended toward the southeast to include the Manitoba Escarpment in the Duck Mountain and Riding Mountain areas (Bamburak, 1999).

Basement Reactivation

Reactivation of the Precambrian basement along the SBZ has been discussed by numerous authors and was reviewed by Bamburak (1999). The main points of the discussion are as follows:

1) Paleozoic strata straddling the SBZ and CGL were probably differentially affected by basin subsidence and uplift during deposition. The Superior craton (and therefore the CGL) underwent greater subsidence during depositional episodes and greater uplift during erosional episodes than the SBZ.
2) Evidence for movement of the SBZ during the Early Paleozoic has been documented in the William Lake area to the northeast and along the Birdtail–Waskada axis to the south. Continuation of these movements during the Mesozoic, along the SBZ, is suggested by the presence of accretionary lapilli in the Easterville area.

3) Penetration of Paleozoic limestone by deep groundwater flow within the outcrop belt is evidenced by solution chimneys in the Mafeking quarry, by the Ochre Lake structure and by modern salt springs. Grasby (GS-34, this volume) describes the latter in further detail.

CURRENT STUDY

Geophysical Surveys

As stated earlier, the presence of the 30 mGal negative anomaly in the Camperville area was documented by the Dominion Observatory in the mid-1960s (Buck, 1968). The gravity low is located immediately east of the SBZ. To test the validity of this low, a series of reconnaissance and detailed transects were conducted during the summer across the low and into the adjacent SBZ. A two-person reconnaissance gravity survey was conducted, from a base in Winnipegosis, over a three-day period in May. The survey was conducted at a 10 km station spacing along three east-west transects across the CGL from Lake Winnipegosis to the Superior Boundary Zone. (Fig. GS-32-3).

The gravity meter used, on loan from the Department of Geological Sciences, University of Manitoba, was a Worden gravity meter with a scale constant of 0.46 mGal/division. The readings could be taken to an accuracy of 0.1 scale division. Elevations were recorded using a GPS unit, but stations were also selected where contour lines on a 1:50 000 scale topographic map crossed the highways. Station spacings for the gravity survey were kept within 10 km. In total, 45 stations were measured along the transects. The station interval of the Dominion Observatory survey was similar to that of the present survey.

The area has little topographic relief except in the vicinity of Duck Mountain. Therefore, only a few stations required topographic corrections. The resulting gravity anomaly of the present survey is identical to that of the 1965 survey.

Two University of Manitoba undergraduate students under our direction carried out a detailed gravity and magnetic transect with 1 km station intervals (*see* Surasky and Minkus, GS-33, this volume). The transect ran from the town of Camperville along PTH 20, PTH 10 and Provincial Road (PR) 275 to the Saskatchewan border (Fig. GS-32-3). Flagged wooden pickets were placed at each station. They used the same gravity meter and, for part of the survey, the University's magnetometer. Unfortunately, their magnetometer malfunctioned midway through the survey and only gravity readings were taken at the remaining stations. We completed the survey in August using the Department's magnetometer, a proton precession MP3 with a sensitivity of 1 nT. The pickets placed by the students enabled us to take measurements at the same locations as the gravity readings. The magnetic readings of the overlapping stations from both surveys were identical.

Gravity readings were also taken along a short transect (8 stations with a 0.5 km spacing) along an east-west township road approximately 5 km north of PTH 10 in the Renwer area (Fig. GS-32-3). This work was carried out to determine the strike of the sharp change in gravity readings along the PTH 10 transect. Interpretation of the data is presently being carried out and will be available at a later date.

Geological Investigations

Devonian and Cretaceous bedrock exposures were sampled to determine if unique mineralization occurs in Paleozoic carbonate rocks and Mesozoic shale within the CGL, and if this mineralization has economic potential. The Devonian outcrops are situated mainly along Lake Winnipegosis and the Cretaceous outcrops are found along the Duck Mountain portion of the Manitoba Escarpment (Fig. GS-32-1). Norris et al. (1982) described most the Devonian exposures and McNeil and Caldwell (1981) have documented many of the Cretaceous sections in detail.

Figure GS-32-4 shows the values of Cu, Zn, Pb, Mn and Fe from hole M-6-80, drilled at a quarry on Pine River Road, SE1-5-33-19W (McCabe, 1980). As described above, honey-coloured sphalerite, with minor pyrite, was identified between 110.8 and 113.2 m (Gale and Conley, 2000), in the Winnipegosis Formation transition zone. Immediately above this zone, polymict collapse breccia was identified at the base of the Dawson Bay Formation. Partly bituminous laminated dolomite was found immediately below the zone. Examination of Figure GS-32-4 indicates that, in hole M-6-80:

1) The Ashern Formation is enriched in nickel and iron relative to the overlying Winnipegosis Formation.
2) The uppermost beds of the Winnipegosis Formation are enriched in Cu, Zn, Pb and Ni, but show a reduction in Fe and Mn relative to the underlying beds. This is also true, with the exception of Ni, for the overlying Dawson Bay Formation.
3) Values for Pb, Ni, Fe and Mn are generally high throughout the Dawson Bay Formation, relative the other formations.
4) Values for Pb, Ni, Fe and Mn seem to be slightly higher near the base of the Point Wilkins Member of the Souris River Formation.
5) There are solitary peaks for Zn, Pb and Ni in the upper portion of the Sagamace Member of the Souris River Formation. The values for Mn seem to be increasing upward through the sequence.
6) Plots of Ni and Fe show similar trends throughout the core, although the relative magnitudes are significantly different. Plots of Pb and Mn also show some similarity to those of Ni and Fe, and the Cu and Zn plots also show some correspondence.

The Pine River Junction corehole M-5-00 (Table GS-32-2) was drilled through 121.0 m of Devonian stratigraphy to verify the presence of sphalerite originally encountered in M-6-80. In corehole M-5-00, black mineralization was encountered at almost the same stratigraphic level as in corehole M-6-80, and appears to be the 'blackjack' form of sphalerite. No honey-coloured sphalerite was found in M-5-00.

B-Horizon Soil Sampling

In October, b-horizon soil samples were collected for enzyme leach and geochemical analysis along PTH 10, in the vicinity of the SBZ and CGL boundary (Fig. GS-32-3).

ACKNOWLEDGMENTS

Ian Ferguson of the Department of Geological Sciences at the University of Manitoba is gratefully acknowledged for facilitating the detailed gravity-magnetometer survey, partially funded under a Natural Sciences and Engineering Research Council of Canada (NSERC) grant. Thanks are also extended to Doug Berk and staff of the Midland Core Facility for drilling M-5-00 during the summer of 2000, and for preparing outcrop and core samples for chemical analysis.

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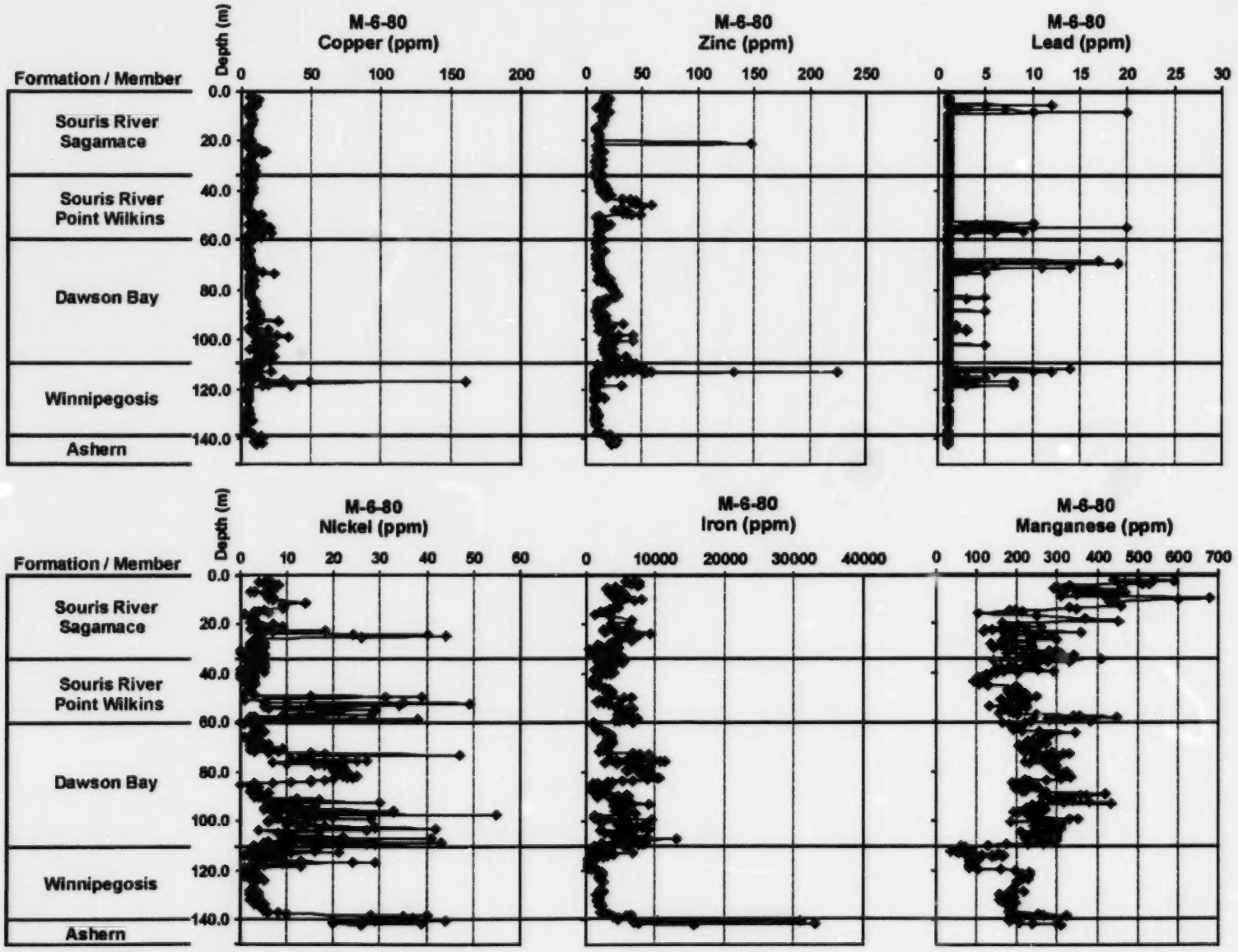

Figure GS-32-4: Concentrations of Cu, Zn, Pb, Ni, Fe and Mn (ppm) in corehole M-6-80.

Table GS-32-2: Detailed log of corehole M-5-00

Hole no.	Location and elevation (m)	SYSTEM/Formation/ (Member)	Interval (m)	Lithology summary
M-05-00	SE1-5-33-19W	DEVONIAN/Souris River	0.0–20.2	Dolomite, fine grained
Pine River	423445N	(Sagamace)		becoming brecciated
Junction quarry	5739324E			downward
	262m		20.2–27.1	Shale, dolomitic shale, limestone, breccia, short intervals of blue clay infill
		(Point Wilkins)	27.1–45.9	Dolomite, breccia, calcite-filled vugs and healed fractures, becoming vuggy below 36.5 m
			45.9–56.6	Shale breccia with disturbed bedding, limestone, dolomite (First Red Beds)
		DEVONIAN/Dawson Bay	56.6–69.8	Limestone, partly dolomitized, fossiliferous, calcite-filled vugs and healed fractures
			69.8–82.8	Calcareous shale, greenish grey becoming purplish red below 71.6 m, mottled
			82.8–94.8	Limestone, brachiopod, calcite-healed fractures, minor pyrite
			94.8–106.7	Polymict collapse breccia; shale, limestone, dolomite; includes Second Red Beds; below 103.0 m, clasts are beige, grey, pink and brown; below 105.0 m, argillaceous greyish green soft clay; at base are carbonaceous stylolites
		DEVONIAN/Winnipegosis (Transition zone)	106.7–112.7	Upper 2.2 m white, ?kaolinitic clay; dolomite breccia, porous, fossiliferous, partly laminated; limestone, fine vuggy, trace pyrite; lowest 0.5 m dark clay laminae with increase in sulphide banding and clots
		(Upper Member)	112.7–117.1	Dolomite, partly bituminous; minor shiny black mineralization (?sphalerite) in fractures at 114.2 m, minor pyrite; below 114.6 m, nine 3 cm thick sulphide-rich laminae and numerous thinner laminae, partly fine vuggy (birdseye) porosity, inter-reef facies
		(Lower Member)	117.1–121.1	Platform facies; dolomite, vuggy, mottled, nodular; massive near top, pinhole porosity near middle and cavernous porosity near base

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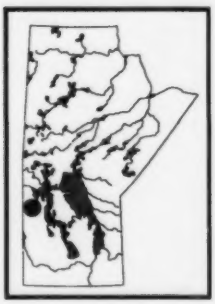
GS-33

GRAVITY AND MAGNETIC PROFILE FROM CAMPERVILLE TO THE SASKATCHEWAN BORDER (PARTS OF NTS 62N/E AND 63C/E)

by A.T. Surasky and R.S. Minkus[1]



Surasky, A.T. and Minkus, R.S. 2000: Gravity and magnetic profile from Camperville to the Saskatchewan border (parts of NTS 62N/E and 63C/E); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 210-213.

SUMMARY

In May 2000, a reconnaissance gravity survey was carried out in the vicinity of the Camperville gravity low in west-central Manitoba (Bamburak et al., GS-32, this volume). The purpose of this survey was to verify the presence of a gravity low in the area. In the present study, a more detailed gravity and magnetic survey was carried out along a line from Camperville to the Saskatchewan border, transecting the area of the gravity low.

INTRODUCTION

Gravity and magnetic surveys were completed during a five-day period, starting June 8, in the vicinity of the Camperville gravity low in west-central Manitoba. The purpose was to obtain a more detailed view of the gravity and magnetic field in the area. The surveys were carried out along Provincial Trunk Highway (PTH) 20, PTH 10 and Provincial Road (PR) 275, from Camperville to the Saskatchewan border. The survey area is shown in Figure GS-33-1.

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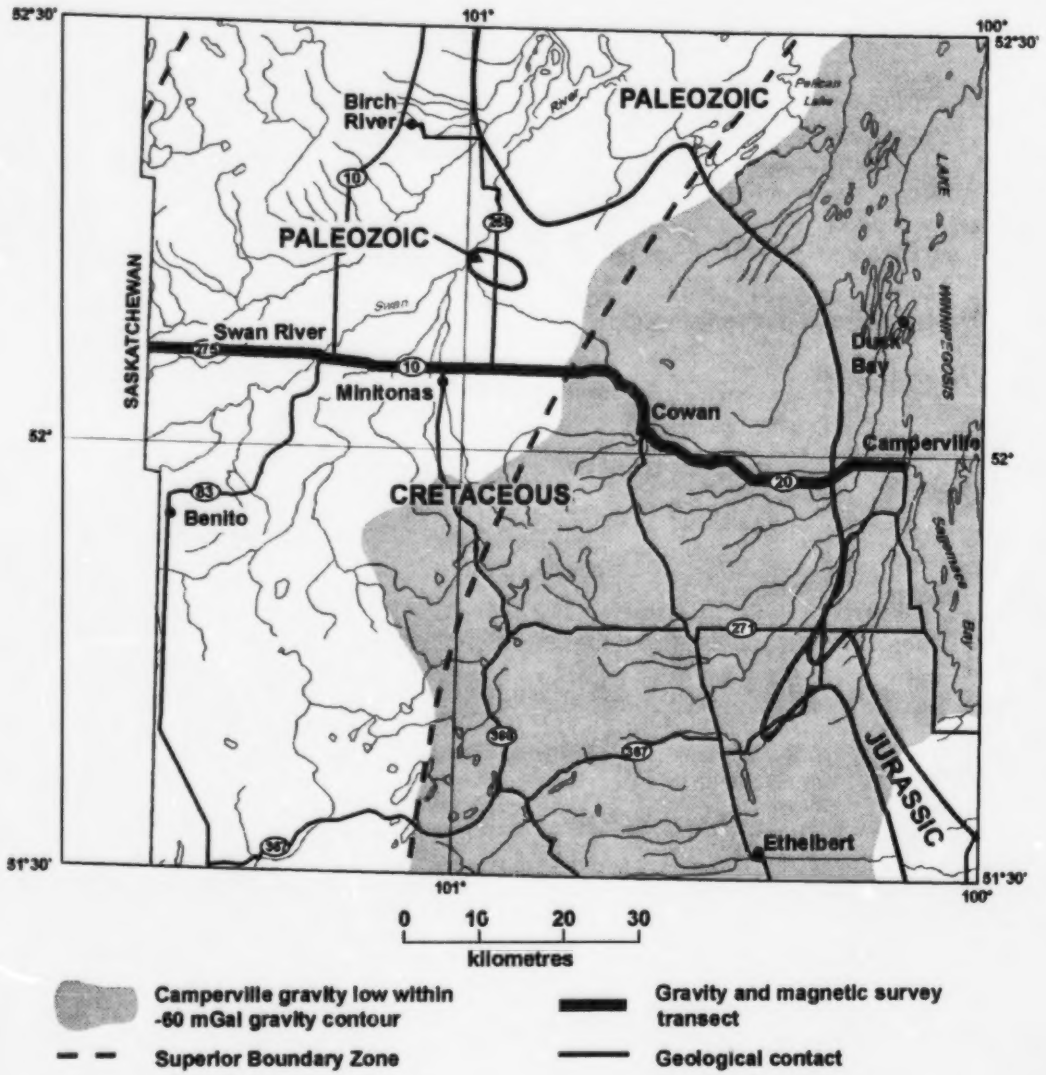

Figure GS-33-1: Map of survey area.

GEOPHYSICAL SURVEYS

Gravity Survey

The gravity survey was completed using a Worden™ gravity meter (model III, no. 543, built by Texas Instruments), with a sensitivity of 0.04 mGal, belonging to the Department of Geological Sciences, University of Manitoba. Readings were taken in the ditch, adjacent to the highways, at 1 km intervals. The gravity meter was situated on a tripod approximately 50 cm off the ground when the readings were taken. The gravity survey used a base station in order to correct for drift. The base station was visited approximately every two hours.

The site position was recorded using a Garmin 12 CX GPS system. As the elevation data from the GPS system are not sufficiently accurate for gravity surveying, station heights relative to the centre of the highway were recorded.

The elevation data were obtained using highway profile maps from the Manitoba Department of Highways and Government Services. Since the highway profiles contain the elevation data for the highways, the station was located on the highway profile and the elevation was read off the profile. The difference in height between the highway centreline and the station was subtracted from the elevation determined from the highway profile, giving the actual elevation of the station.

Data Reductions

Gravity data need to be reduced so that only lateral subsurface density variations are represented by the data (Reynolds, 1997). In the processing of the gravity data, the following corrections were made in order to reduce the data to the Bouguer anomaly:

- Drift correction (using repeat base station readings)
- Free-air correction

  $\delta g_F = 3.086h$ where h is the height differential between stations, and δg_F is in gravity units (g.u.)
- Bouguer correction

  $\delta g_B = 2\pi G\rho h$ where $G = 6.67 \times 10^{-8}$ $m^3 \cdot Mg^{-1} \cdot s^{-2}$, ρ = density (a density of 2.67 $Mg \cdot m^{-3}$ [2670 $kg \cdot m^{-3}$] was assumed), δg_B is in g.u.
- Latitude correction

  $\delta g_L = -8.108\sin(2\phi)$ where ϕ is degrees of latitude, δg_L is in g.u./km

The terrain correction was not applied because the survey area had limited relief and, with the assumed density, the correction was negligible.

Results

The Bouguer gravity anomaly profile is shown in Figure GS-33-2. There is a definite decrease in gravity progressing from west to east. There is a 60 mGal (600 g.u.) change over the length of the survey. There is an increase of approximately 10 mGal (100 g.u.) in the Bouguer anomaly between 350000m east and 375000m east. There are also shorter wavelength (<10 km) and smaller amplitude (<3 mGal) variations toward the eastern end of the survey (Fig. GS-33-3).

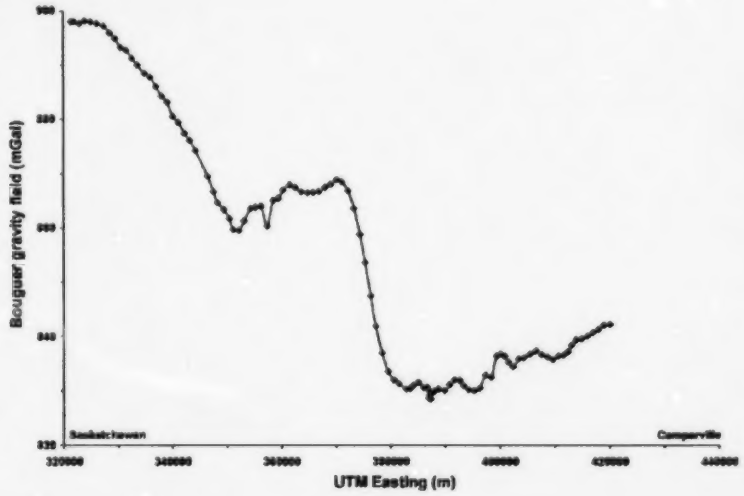

Figure GS-33-2: Profile of the Bouguer gravity anomaly showing a general rise in gravity from east to west.

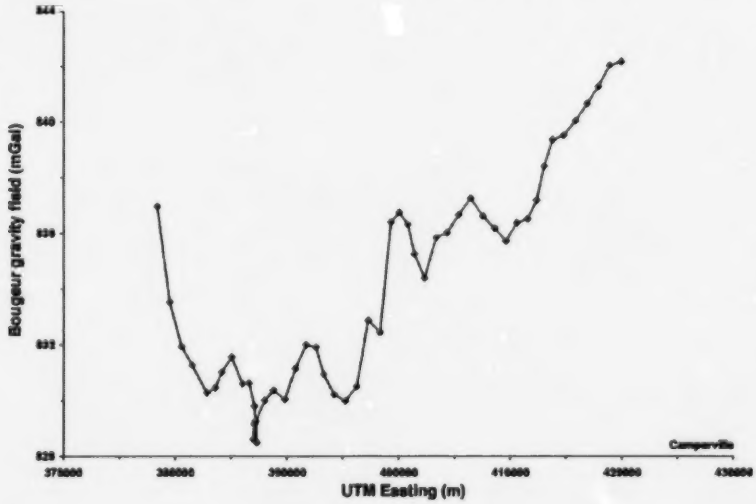

Figure GS-33-3: Enlargement of the Bouguer gravity anomaly for the eastern 50 km of the transect, showing shorter wavelength and smaller amplitude responses.

Modelling

Some preliminary two-dimensional modelling of the gravity data was done using the computer program Potent™, produced by PC Potentials. A constant average value was subtracted from the data prior to modelling. The aim of the modelling was to derive a simple model that fit the data. The model consisted of three polygonal prisms located within the upper crust. The data were fitted by iteratively adjusting the density of the bodies and the apices of the polygons. The final model is shown in Figure GS-33-4. The data were well fitted by the three bodies. The final density contrast of the bodies is +0.17, +0.20 and -0.10 $Mg \cdot m^{-3}$ with respect to the background density. This model is just preliminary but, if the true geological source of the anomalies has a size similar to the polygons, the observed anomalies can be explained by a density contrast on the order of 0.20 $Mg \cdot m^{-3}$.

Magnetic Survey

The magnetic survey was completed using the Scintrex MP-4 proton precession magnetometer and the IGS-2, both with a sensitivity of 1 nT and belonging to the Department of Geological Sciences, University of Manitoba. Total magnetic field readings were taken at the same stations as the gravity measurements. Two total magnetic field readings were taken at each station. A magnetic base station was used, and was revisited approximately every two hours. Due to an instrument malfunction, the magnetic survey could not be completed in June, and was instead completed in late August (Bamburak et al., GS-32, this volume). Repeat measurements at common stations from both surveys showed good correlation between the data.

Data Reductions

The magnetic field drift was removed from the June data by using total magnetic field repeat readings taken at the base station. However, for the data collected in August (stations 49–104), the magnetic diurnal variation was negligible and was not removed.

Results

The total magnetic field profile is shown in Figure GS-33-5, and an enlargement of the eastern 50 km is shown in Figure GS-33-6. There appears to be a general rise in the magnetic field from west to east, but the magnetic response is not as smooth as the gravitational response. Initial comparison of the Bouguer gravity anomaly and total magnetic field suggest an inverse correlation.

Future work will include modelling of the total magnetic field data, and joint modelling of the Bouguer anomaly and the total magnetic field.

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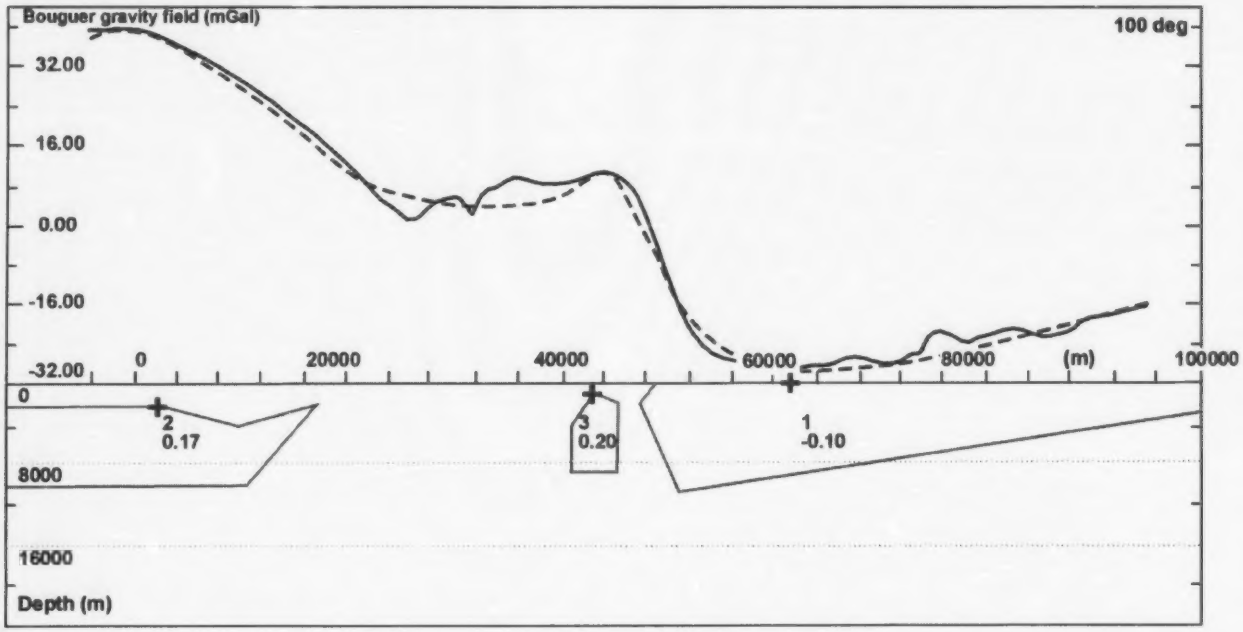

Figure GS-33-4: Preliminary model fitted to the data. The model consists of three upper crustal bodies with density contrasts of +0.17, +0.20 and -0.10 $Mg \cdot m^{-3}$ with respect to the background density.

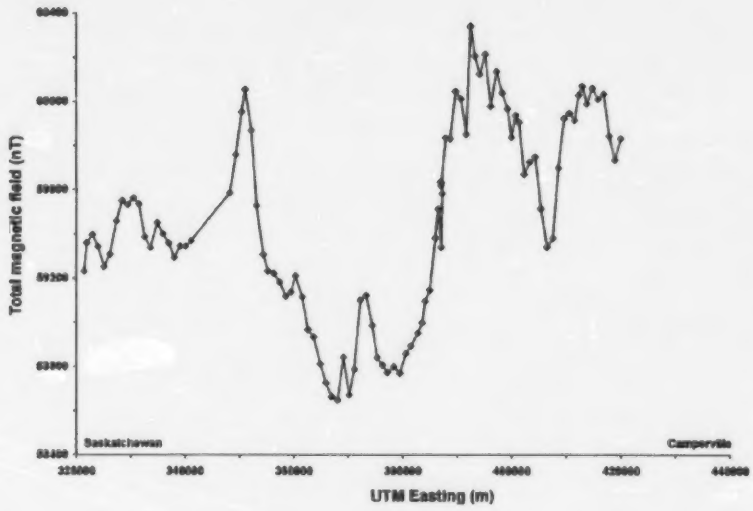

Figure GS-33-5: Profile of total magnetic field response showing an inverse relation to the Bouguer gravity anomaly.

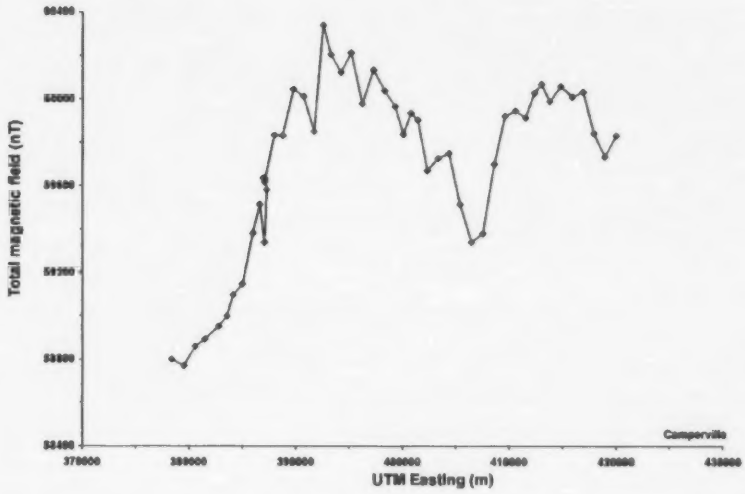

Figure GS-33-6: Enlargement of the total magnetic field response for the eastern 50 km of the transect.

SALINE SPRING GEOCHEMISTRY, WEST-CENTRAL MANITOBA

by S.E. Grasby[1]

Grasby, S.E. 2000: Saline spring geochemistry, west-central Manitoba; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 214-216.

SUMMARY

Saline spring waters have been sampled along the west shore of Lake Winnipegosis. Preliminary results suggest Pleistocene glaciation had a significant impact on the regional flow system of the Paleozoic aquifers in the area.

INTRODUCTION

The carbonate rock aquifer constitutes the fresh water–bearing portion of the Paleozoic outcrop belt in southern Manitoba, and is a principal groundwater resource for Manitoba. Over the last three years, the Geological Survey of Canada, in co-operation with Manitoba Water Resources and the Manitoba Geological Survey, has conducted a regional assessment of the hydrogeology and hydrogeochemistry of the aquifer. The fresh water–bearing portion of the aquifer is bounded on the east by the erosional limits of the Paleozoic outcrop belt, and on the west by a hydrological divide defined by the Rat, Red and Assiniboine rivers, and lakes Manitoba and Winnipegosis. West of this divide, the salinity of waters rapidly increases to that of typical formation waters in the Williston Basin. Potential eastward migration of these saline brines would threaten the long-term stability and quality of the fresh water–bearing portions. The purpose of this study was to conduct a geochemical and stable-isotope analysis of brine waters in order to understand better their origin and how they relate to the regional flow system. The saline springs along the west shore of Lake Winnipegosis offer the easiest means of sampling these waters.

HISTORICAL PERSPECTIVE

The earliest reports of saline springs in Manitoba are from the journals of La Verendrye in the early 1700s. By the early 1800s, there are records of salt-production operations in the area. Although providing an important source of salt for early settlers, these operations had been abandoned by the 1880s due to competition from rock salt being shipped by rail from Ontario (Petch, 1990).

There is little previous work on the saline springs of Manitoba. The first chemical analyses were reported by Tyrrell (1892). Since then, they have been sampled by Stephenson (1973), van Everdingen (1971) and Wadien (1984). However, water chemistry and stable-isotope data from these studies is either not complete, inaccessible or limited to a minor number of springs. Geochemical sampling of precipitates from many of these sites has been conducted by Bezys et al. (1997).

REGIONAL DISTRIBUTION

The regional distribution of saline springs in Manitoba is depicted in Figure GS-34-1. Most sites have been found either through historical

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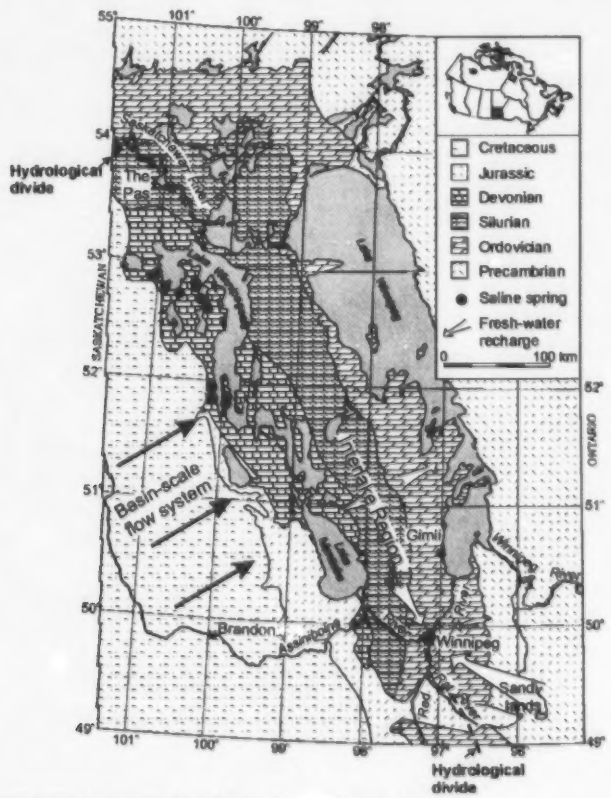

Figure GS-34-1: Regional map showing known locations of saline springs.

records or by use of airphotos. Spring sites are often readily apparent on airphotos because surrounding vegetation has been killed off by the saline water. To date, more than 65 sites have been identified that have currently flowing springs or show signs of recent activity. Forty-five sites have been sampled to date for water chemistry. It is interesting to note that people living in the Dawson Bay area tell stories of portions of Lake Winnipegosis that do not freeze during the winter and trucks falling through areas of weak ice. This suggests that saline waters may also discharge into the lake, thereby locally reducing the freezing temperature of the water.

There is a marked difference in character in discharge sites going from south to north. In the southern sites, springs are surrounded by grassy meadows or light brown to grey mud. In contrast, springs in the northern part tend to have large amounts of iron staining, forming red mounds around the spring outlet and over an extensive area surrounding it. As well, the northern sites typically have large growths of bacteria mats around the outlets; these are absent in the south.

SPRING CHEMISTRY

Stable-isotope data for the springs are plotted in Figure GS-34-2, along with the global meteoric water line (GMWL). Spring waters tend to plot along the line and are distinctly different from values typical of formation brines observed west of Manitoba (Rostron et al., 1998). This indicates that the saline springs originated as meteoric water. The total dissolved solids (TDS) load of the spring sites is variable and ranges from 2000 to 114 000 mg/l. Springs with the highest TDS are in the Pelican Bay area. Given the meteoric origin of the waters, the high TDS content must then be due to dissolution of evaporite beds. The piper plot in Figure GS-34-3 illustrates that spring waters are all Na-Cl in composition. Figure GS-34-4 shows a good linear correlation between Na and Cl, consistent with dissolution of halite. The variation in concentration is likely a complex relationship involving residence time in the subsurface (dictating the amount of dissolution and water/rock interaction that occurs) and the degree of mixing with shallow groundwaters.

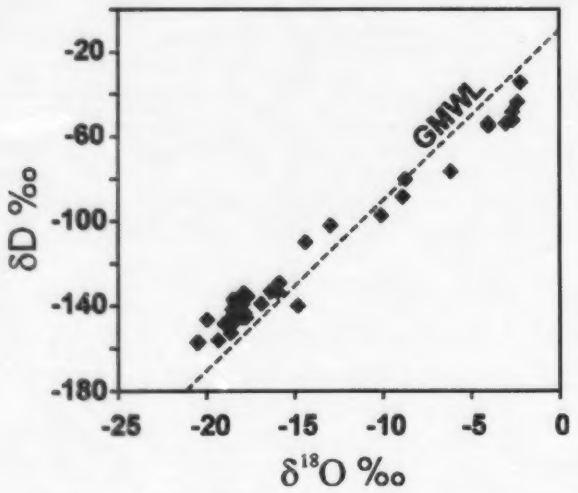

Figure GS-34-2: Plot of δD versus $\delta^{18}O$ for saline spring waters. GMWL, global meteoric water line.

DISCUSSION

An important characteristic of saline spring waters is that they have a stable-isotope signature consistent with meteoric water. This implies that there must have been a large flux of fresh water into the aquifer. Previous models have either interpreted this as local influx of groundwater (e.g. Hitchon, 1969), or basin-scale flow of meteoric waters, from recharge sites in South Dakota, around the edges of basin brines in the deepest portions of the Williston Basin to discharge sites in Manitoba (e.g. Bachu and Hitchon, 1996). The first model invokes crossformational flow, which tends to have inherently lower permeability than flow along bedding. The second model involves long-range transport of fluids with little or no mixing. As an alternative, Grasby et al. (2000) have suggested that these saline meteoric waters represent a significant influx of glacial meltwater during the Pleistocene. The hydrological head due to overlying ice sheets would have been large enough to drive basal meltwater into the Paleozoic successions that outcrop in southern Manitoba. Grasby et al. (2000) estimated that metlwaters may have reached as far as 300 km into the Williston Basin. Here, the fresh waters would readily dissolved evaporite beds. The retreat of ice sheets would allow the re-establishment of the present southwest to northeast flow system of the Williston Basin (Bachu and Hitchon, 1996), thus pushing the meltwaters back to surface, where they express themselves as the saline springs observed today.

Future work on modelling water-rock interaction and regional hydrodynamics will test this hypothesis.

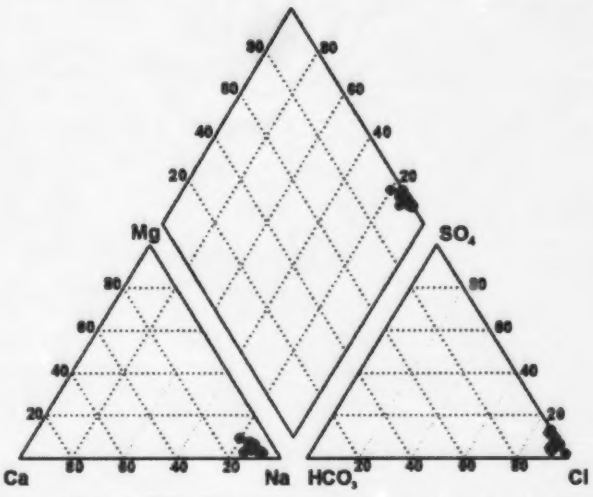

Figure GS-34-3: Piper diagram (in meq/l) showing major-ion composition of spring waters.

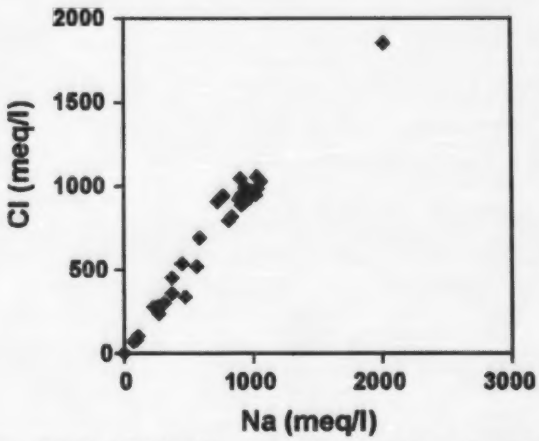

Figure GS-34-4: Plot of Na versus Cl, illustrating a strong linear correlation.

ACKNOWLEDGMENTS

Ruth Bezys is gratefully acknowledged for providing locations of springs and information on access. Kathleen Bradford is thanked for cheerfully lugging sampling equipment through numerous swamps.

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GS-35 AGGREGATE RESOURCES IN THE RURAL MUNICIPALITY OF CAMERON

by H. D. Groom

Groom, H.D. 2000: Aggregate resources in the Rural Municipality of Cameron; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 217-219.

SUMMARY

The Rural Municipality (R.M.) of Cameron covers eight townships in southern Manitoba. Most of the municipality was inundated by glacial Lake Hind and much of the surface material is lacustrine sand. Gravel reserves are in two delta deposits. These deposits consist of 3 m to more than 6 m of sand and gravel, with most of the reserves being below the watertable. Shale content varies quite widely within the deposits.

There are six active gravel pits and three shale quarries in the area. Most of the aggregate produced is for use within the municipality. The municipality is the greatest consumer of aggregate, mostly for road maintenance.

INTRODUCTION

The R.M. of Cameron recently joined with the R.M.s of Pipestone and Sifton to form the Denis County Planning District. The aggregate resources in the R.M.s of Pipestone and Sifton were mapped previously (Groom, 1994; UMA, 1977). The aggregate deposits in the R.M. of Cameron, however, had never been mapped at a scale suitable for submission to the district planning board. A sand and gravel inventory was conducted this summer in order to provide aggregate information at 1:50 000 scale for resource management and land-use planning.

LOCATION AND PHYSIOGRAPHY

The municipality covers eight townships (twp. 5 and 6, rge. 22 to 25W) in southwestern Manitoba. It is primarily a farming district and the town of Hartney is the major service centre. The Canadian Pacific Railway, Highway 21 and a network of gravel roads are the major transportation routes. Souris River runs northeastward through the centre of the municipality.

The area is divided into two physiographic regions: the Boissevain plain in the southeastern third of the municipality and the Souris basin in the remainder of the area.

Surficial sediments on the Boissevain plain have an average thickness of 5 m (Betcher, 1983). The plain has a gently rolling topography imparted by linear till ridges, except in the area of the Dand channel. The channel is up to 0.5 km. wide, 10 to 15 m deep and trends northeast for approximately 20 km before ending at a gravel deposit at the eastern edge of the municipality.

Surficial sediments in the municipality thicken westward, attaining thicknesses of more than 70 m in the northwestern part of the area (Klassen and Wyder, 1970). The sediments are primarily lacustrine silt and clay south of the river, and sand in the western and northwestern parts of the municipality. In the area of the Lauder Sand Hills, the sand has been blown into spectacular dunes that are now stabilized by vegetation. Deltaic gravel deposits are sparsely scattered throughout the area.

BEDROCK GEOLOGY

The municipality is underlain by the Odanah Member of the Pierre Shale Formation. The Odanah Member is a hard, grey, siliceous marine shale with thin interbeds of soft, olive-grey shale. Bentonite beds occur throughout the unit. Bannatyne (1970) assigned this unit to the Riding Mountain Formation, but work by McNeil and Caldwell (1981) resulted in a revised nomenclature that abandoned the use of the term 'Riding Mountain Formation' and assigned its members to the Pierre Shale Formation.

Bedrock topography maps show the gentle swale of the buried Pierson valley, an early Pleistocene channel eroded into the bedrock (Klassen and Wyder, 1970). Glacial sediments thicken westward as they infill the valley. For this reason, bedrock is rarely exposed in the municipality. It does outcrop in the Dand channel where glacial sediments have been eroded away. The quarry in SW18-5-22W is situated on the east edge of the channel and provides a good exposure of the Odanah shale.

PREVIOUS WORK

The bedrock geology was mapped by Wickenden (1945) and Bannatyne (1970), and the units were further described by McNeil and Caldwell (1981). A map of the regional bedrock topography, at a scale of 1:500 000, has been produced by Klassen et al. (1970). Klassen and Wyder (1970) discussed the location and fill of buried valleys; the bedrock topography map that accompanies their report is at a scale of 1:250 000.

Maps of drift thickness are included in groundwater studies prepared by Betcher (1983).

Elson (1956, 1961) mapped the surficial geology of the area at a scale of 1:126 720 and outlined the glacial history as part of a regional study of southwestern Manitoba and southeastern Saskatchewan. A portion of this area, the east half of the Virden map sheet (NTS 62F), has recently been remapped at a scale of 1:100 000 (Sun and Fulton, 1995) as part of the Prairie NATMAP Project of the Geological Survey of Canada. The R.M. of Cameron falls within this map area.

Sun (1996) reconstructed the late glacial history of the area, with special emphasis on the levels of glacial Lake Hind over the course of its existence. Much recent work has focused on the effects of catastrophic flooding as glacial lakes in eastern Saskatchewan burst their ice dams (Kehew, 1982; Kehew and Clayton, 1983; Kehew and Teller, 1994). Several of the deltas in Lake Hind are related to flood waters originating in Saskatchewan (Sun, 1996; Sun and Teller, 1997).

LATE GLACIAL HISTORY

A generalized map of the surficial sediments in the region is shown in Figure GS-35-1. The sequence of late Wisconsinan events, particularly the deglacial history of the area, is outlined in Elson (1956) and Sun (1996). The following brief history of events is taken primarily from Sun (1996) and Sun and Teller (1997).

The surface till was deposited by the southeast-flowing Assiniboine Lobe during the Late Wisconsinan. The Red River Lobe flowed to a position just east of the R.M. of Cameron. These two ice masses blocked the regional drainage and caused glacial lakes to form at the ice fronts: glacial Lake Souris formed at the south margin of the Assiniboine Lobe and a precursor to glacial Lake Hind ponded between the Red River Lobe and the north margin of Assiniboine Lobe. The Dand channel was cut when water from glacial Lake Souris flowed overland to this lake. The deltaic deposit in twp. 6, rge. 22 was formed at this time. As the Assiniboine Lobe retreated, the two lakes merged. Falling water levels caused the emergence of a height of land near Melita that separated the lake into two basins: glacial Lake Hind in Manitoba and the main body of glacial Lake Souris in North Dakota. Several deltas formed along the west edge of glacial Lake Hind during this period, but none occur in the R.M. of Cameron. Two catastrophic floods from glacial lakes in Saskatchewan affected glacial Lake Hind. The first deposited the delta at

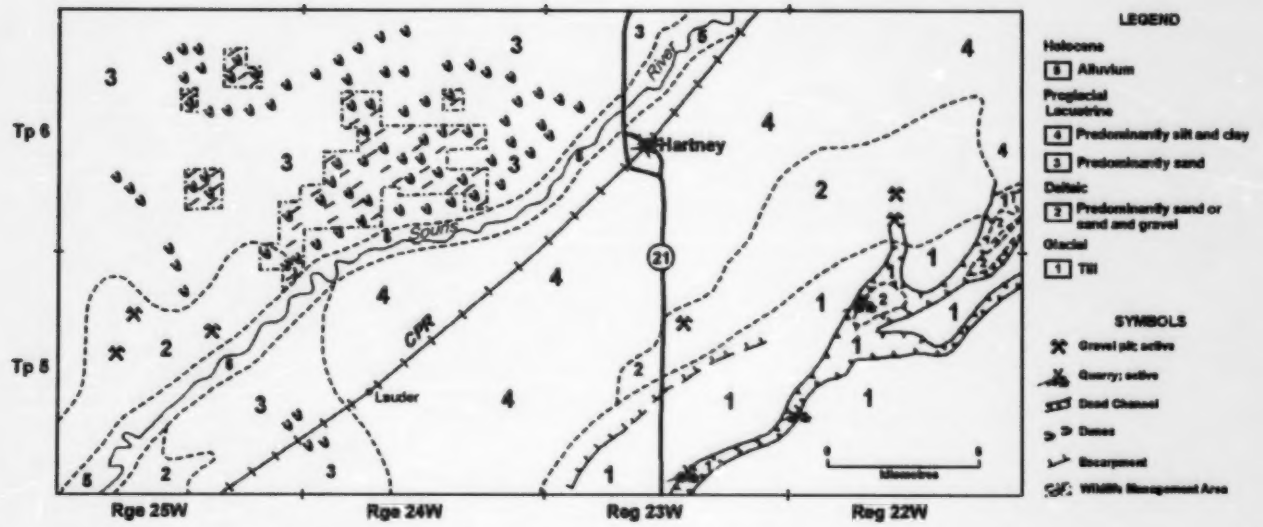

Figure GS-35-1: Surficial geology of the Rural Municipality of Cameron (after Sun and Fulton, 1995).

Melita outside the municipality and the second created the deltaic deposit west of Lauder in twp. 5, rge. 25 in the R.M. of Cameron. Later floods along the Assiniboine River valley deepened the Pembina trench, resulting in final drainage of glacial Lake Hind.

METHODOLOGY

Surficial deposits were delineated on 1:20 000 scale township photomosaics. Deposit delineation was based on surficial units outlined by Sun and Fulton (1995). The township mosaics were also used as a base on which to compile aggregate information from several sources:

- Active pit locations – Mines Branch quarry database
- Quarry lease and withdrawn locations – Mines Branch plat books
- Crown versus private ownership – Manitoba Crown Lands Branch database
- Pit and sample locations – Department of Highways block files
- Water-well records from Manitoba Conservation (GWDrill)

Gravel pits, roadcuts and natural exposures were examined during the first part of the field examination. Pits were examined for type of material, degree of depletion and active/inactive/depleted status. Unopened portions of deposits were inspected and land uses that would limit aggregate extraction noted.

Aggregate samples were processed in two stages. In the field, samples that weighed between 75 and 100 kg were passed through 6-inch (15.2 cm), 3-inch (7.5 cm), 1fi-inch (3.8 cm) and fl-inch (1.9 cm) screens. The weights of the 3-, 1fi-, fl- and less than fl-inch fractions were recorded. In addition, the relative abundance of less than 6-inch material, as well as deleterious material (e.g. shale, concretions), was noted. Pebble counts on the 1fi- to 3-inch fraction were done in the field. A representative sample of the less than fl-inch fraction was taken for processing by the Material and Research Branch of the Department of Highways, under the terms of the cooperative agreement initiated in 1998.

The second stage is a backhoe program designed to test the extent, depth and quality of deposits identified earlier. Eight quarter-sections were scheduled to be tested during the last week of September.

AGGREGATE RESOURCES

Much of the municipality, especially north of Souris River, is underlain by lacustrine sand deposited into glacial Lake Hind. The sand ranges from coarse to fine and, in many places, has been blown into dunes. The sand has little economic value and, despite the abundance of sand, aggregate is actually scarce. There is little crown land in the area and most of the aggregate, including shale bedrock, is privately owned.

There are six active gravel pits and three shale quarries. The municipality is the largest user of aggregate, owning two pits and operating in several others. Most of the aggregate produced in the municipality is for municipal consumption.

Shale

The three shale quarries are all associated with the Dand channel (Fig. GS-35-1). Two of them are actual quarries, being excavated out of Odanah shale bedrock. These are roadside excavations, with 1 to 5 m of till overburden. The third is technically a gravel deposit. The material is 99% shale cobbles with medium sand fill. The material from all three sites is used by the municipality.

Sand and Gravel

Gravel in the municipality is located in two deltaic deposits: the Dand delta in rge. 22 and 23 and the Lauder delta in rge. 25 (names taken from Sun, 1996).

Water-well records show that the Dand deposit consists of 3 to 6 m of sand and gravel over till. In pit exposures, the material ranges from fine pebble gravel to cobble gravel. Shale content varies but seems to be higher in the eastern part of the delta. The deposit has three active pits. One (NE3-6-22W) is new but the others have been in use for many years. Most of the remaining reserves are below the water table.

The Lauder deposit has more than 6 m of sand and gravel, once again with most of the reserves below the water table. Two of the active pits are long established and both are water floored. The municipal pit in 20-5-25W has a reported 4 m of aggregate below the water table. Extraction is normally by dragline, but this year the pit was being expanded southward above the waterline. Shale content is low in this pit. The active pit in 22-5-25W seems to have coarser material and more shale, but good exposure in this pit was very limited. The crown gravel in NE20-5-25W has been extensively mined. Two of the pits have recently been rehabilitated and dune sand covers much of the remaining area.

CONCLUSIONS

The municipality utilizes both bedrock and gravel for its aggregate needs. The two delta deposits are a good source of aggregate. However, most of the remaining reserves are below the water table. The added cost of production and the shale content in parts of the deposits decrease their economic value.

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PALEOENVIRONMENTAL HISTORY OF THE RED RIVER VALLEY SINCE AD 1463

by E. Nielsen and S. St. George[1]

Nielsen, E. and St. George, S. 2000: Paleoenvironmental history of the Red River valley since AD 1463; *in* Report of Activities 2000; Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 220-222.

SUMMARY

The Manitoba Geological Survey and the Geological Survey of Canada have initiated a large, multidisciplinary research program to reconstruct the pre-instrumental flood history of the Red River, and to assess the importance of geological processes that may be increasing the flood hazard. The project is using a variety of geological and biological records to provide annual reconstructions of important hydroclimatic variables over the last several thousand years and provide a proxy record of high-magnitude floods in the Red River valley. The current tree-ring record was collected over a 100 km transect along the Red River and extends for 536 years from AD 1463 to 1999. This record contains anatomical signatures caused by Red River floods in 1950, 1852, 1826, 1811, 1778, 1747 and 1532. Complementary isotopic, chemical and ringwidth records will be developed to evaluate the relationship between climatic conditions within the Red River drainage basin and the flood record of the Red River.

INTRODUCTION

Flooding of the Red River is a significant recurring natural hazard in southern Manitoba. About 70 per cent of Manitoba's 1.1 million citizens live in the Red River valley, making the province particularly vulnerable to social and economic disruption during widespread flooding. During the so-called 'flood of the century' in 1997, costs for flood fighting and recovery in Manitoba were estimated at $500 million and 28 000 people were evacuated from high flood-risk areas. Without a series of measures protecting Winnipeg, losses due to flooding could have reached $5 to 7 billion.

Instrumental flood records for the Red River begin in 1912 but have been extended using historical records to cover most of the 19th century. In response to the 1997 Red River flood, the Manitoba Geological Survey (MGS) and the Geological Survey of Canada (GSC) have initiated a large, multidisciplinary research program to reconstruct the pre-instrumental flood history of the Red River, and to assess the importance of geological processes that may be increasing the flood hazard. A paleoflood record will increase the understanding of the recurrence interval of extreme floods (those having a magnitude equal to or larger than the flood of 1826). This record will aid decision-making about, for example, whether the capacity of the Red River floodway needs to be increased. A variety of proxy flood records will be developed from the analysis of Lake Winnipeg cores by L.H. Thorleifson (GSC–Ottawa) and M. Lewis (GSC–Atlantic), of Red River alluvium by G. Brooks (GSC–Ottawa) and of tree rings by S. St. George (GSC–Winnipeg) and E. Nielsen (MGS). These data will provide the primary information for determining how the frequency and magnitude of Red River floods have responded to long-term changes in climatological and environmental conditions over the past several centuries.

The tree-ring component of the project aims to:

1) provide annual reconstructions of important hydroclimatic variables within the Red River valley over the last 350–400 years;
2) provide a proxy record of high-magnitude floods of the Red River; and
3) place the flood of 1826 within a context of approximately 500 years.

SAMPLE COLLECTION

Since fall 1999, tree-ring samples were collected from 12 live oak sites along the Red River to supplement the six sites sampled previously (Table GS-36-1). Mild winter conditions in 1999 allowed sampling to continue until late December. The collection of subfossil oaks preserved in the alluvial sediment of the Red River continued but was limited by high water levels.

Table GS-36-1. Locations and characteristics of bur oak tree-ring sample sites.

Site	Type	No. of trees	Span
Birds Hill Park	Living		NA
Hyland Park	Living		NA
Kildonan Park	Living		NA
Winnipeg	Living	20	1822-1994
	Historical	14	1644-1872
Assiniboine Park	Living		NA
Bruce Park	Living		NA
Monson Park	Living		NA
Barber House	Historical	35	1648-1864
St. Vital Park	Living	12	1830-1998
St. Norbert	Living	10	1855-1998
LaBarriere Park	Living	11	1892-1998
Shay	Living	14	1907-1999
Rat River House	Historical	6	1659-1859
Ste. Agathe	Living	11	1866-1998
Parker Farm	Living	12	1884-1998
Morris	Living	3	1874-1988
Horseshoe Lake	Living	15	1907-1999
St. Jean Baptiste	Living	10	1883-1997
Remus Farm	Living	10	1879-1998
Marais River	Living	11	1850-1998
Fort Dufferin	Living	9	1866-1997
	Historical	2	1723-1872
Red River alluvium	Subfossil	25	1463-1997

NA, not available

ANALYSIS

Anatomical Flood Signatures

Dendrochronological analysis of bur oak (*Quercus macrocarpa*) growing along the Red River has developed a continuous tree-ring record that extends for 536 years from AD 1463 to 1999. Two hundred

[1] Geological Survey of Canada, 360-1395 Ellice Avenue, Winnipeg, Manitoba, R3G 3P2

and thirty tree-ring samples were taken from living trees, historical buildings and alluvial deposits in the Canadian portion of the Red River valley. The sampling locations extend more than 100 km along the Red River from Emerson to Birds Hill Provincial Park. The collection of samples from historic buildings and living trees maintains good sample depth back to 1650, but the sample depth from 1463 to 1650 is relatively poor, consisting of a limited number of subfossil logs.

Detailed examination of the specimens has revealed that thirteen trees from four sites along the river contain isolated annual rings with distinctive anatomical anomalies (Fig. GS-36-1; St. George and Nielsen, 2000). These anomalies are characterized by a marked reduction in the size of earlywood vessels for the rings formed in the years 1950, 1852, 1826, 1811, 1778, 1747 and 1532. While the rings from 1950 document the influence of the well-known flood of that year, the 1852 and 1826 rings correspond to the occurrence of the second largest and largest, well documented Red River floods, respectively. The anomalous rings are interpreted to be the result of inundation of the trees by floodwaters during the spring and early summer. By extension, the anatomical anomalies in 1811, 1778, 1747 and 1532 are interpreted as signatures of major, and previously unknown, floods of the Red River.

The results so far have yielded extremely encouraging results regarding the identification of past major floods along the Red River. The collection and analysis of tree samples and the identification of rings with anatomical anomalies are ongoing and work will continue into 2000–2001.

Precipitation Reconstruction

The incidence of heavy-precipitation events, thunderstorms and cyclones in north-central North America has been on the increase since 1920 (Changnon and Kunkel, 1995). However, most instrumental records are too short to provide an accurate assessment of the relationship between changing climates and flood frequencies. Studies in western North America have demonstrated that long records of precipitation may be derived from long-lived, climatically sensitive trees growing at forest-prairie ecotones (e.g. Case and MacDonald, 1995; Woodhouse and Meko, 1997). Up to this point, however, dendroclimatic reconstructions in Canada have largely been restricted to alpine sites in British Columbia and western Alberta (Case and MacDonald, 1995; Luckman et al., 1997; Watson and Luckman, in press; St. George and Luckman, in prep.). Although these reconstructions provide high-quality, high-resolution records of past climate, they cannot be considered representative of conditions experienced in the central and eastern Prairies. This is particularly true for reconstructions of precipitation, as drought has been regionally asynchronous in North America during the 20th century (Woodhouse and Overpeck, 1998; MacDonald and Case, 2000). While the relative scarcity of long-lived trees within the Canadian Prairies has been a significant obstacle to developing local dendroclimatic reconstructions, the recovery of well-preserved subfossil trees from the alluvium of prairie rivers holds great promise for developing long, moisture-sensitive, tree-ring chronologies.

Preliminary analysis suggests that bur oak ringwidth is highly correlated with annual precipitation in the Red River valley during the 20th century (St. George and Nielsen, in prep.). Therefore, the oak network should provide an accurate reconstruction of hydrological conditions in southern Manitoba since the middle of the 15th century.

Isotopes

Work has been initiated to determine changes in $\delta^{13}C$, $\delta^{18}O$ and $\delta^{2}H$ isotope levels over the last 350 years for *Quercus macrocarpa* tree-ring specimens collected in Winnipeg. These analyses, which are being

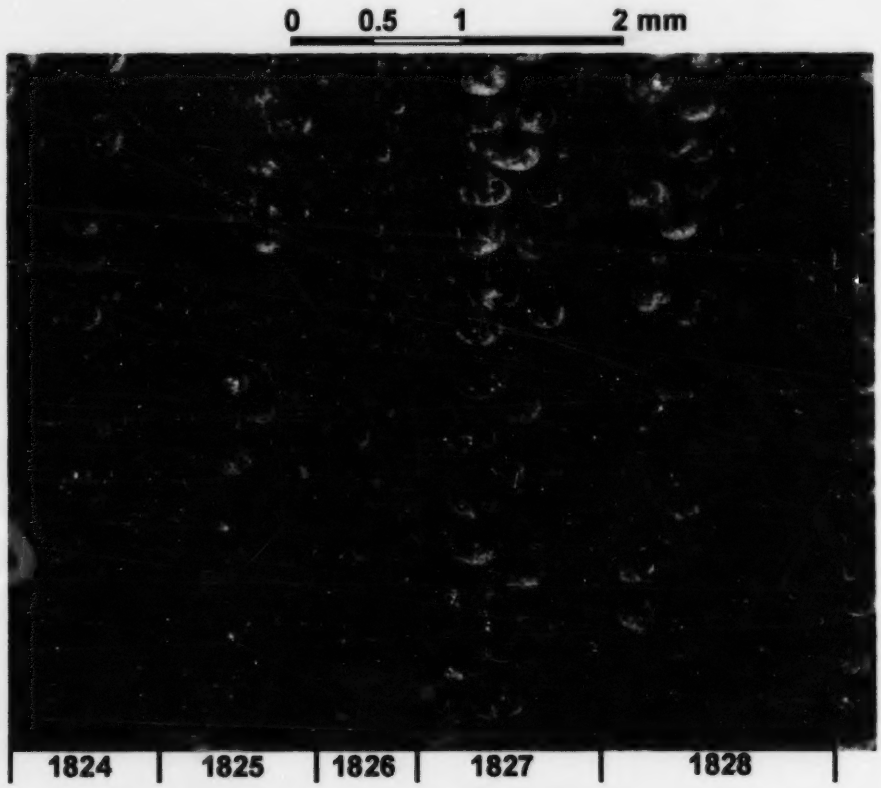

Figure GS-36-1: Flood ring from 1826 in bur oak (Quercus macrocarpa) from central Winnipeg. Vessels appear as light-coloured circles in the left part of each annual ring. The flood ring is distinguished by the reduced area of its earlywood vessels.

done in collaboration with W. Buhay (Department of Geography, University of Winnipeg), will provide a record of climate change independent from that reconstructed from ringwidth. Preliminary results suggest that $\delta^{18}O$ in Red River oaks may be related to variations in summer precipitation.

Chemistry

Analysis of the elemental composition of tree rings is being carried out in conjunction with P. Outridge (GSC-Ottawa), with the aim of identifying chemical flood signatures in tree-rings. These analyses will also attempt to identify anthropogenic influences in the Red River basin. Laser-ablation ICP-MS has previously been widely applied in the geological and materials sciences to retrieve microspatial elemental data from solid materials such as minerals and synthetic products. This technique offers the potential for measurement of more than 50 elements, with sampling spot sizes down to 5 μm diameter.

Our initial investigations have focused on determining whether the technique is sensitive with respect to sampling trace elements in tree rings, and whether trees respond to flood events by laying down rings with significantly perturbed chemistry. Preliminary results suggest that trees from Fort Dufferin contain a dramatic increase in iron levels coincident with the 1997 flood. If the 1997 Red River flood caused this increase, chemical analysis may provide independent confirmation of past flood events in support of the anatomical flood-signature analysis.

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BEDROCK GEOLOGY COMPILATION MAP SERIES

by L. Chackowsky and D. Lindal

Chackowsky, L. and Lindal, D. 2000: Bedrock Geology Compilation Map Series; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 223-224.

SUMMARY

The Bedrock Geology Compilation Map Series (BGCMS) was initiated in 1985 to provide uniform 1:250 000 scale geological compilation maps for the entire province, initially focusing on areas of high mineral endowment and/or high mineral potential. Digitization of all existing manually drafted BGCMS maps is almost complete, and map production is now 100% digital, using ArcInfo geographic information system (GIS) technology. There are now 22 BGCMS maps, covering more than half of the province (Fig. GS-37-1), including the entire Churchill Province, core areas of the Superior Province, and the capitol region. The parts of Manitoba not yet covered by BGCMS include areas of mostly unexposed Phanerozoic bedrock, and parts of the Superior Province with little or no bedrock exposure or supracrustal bedrock (Lenton, 1999).

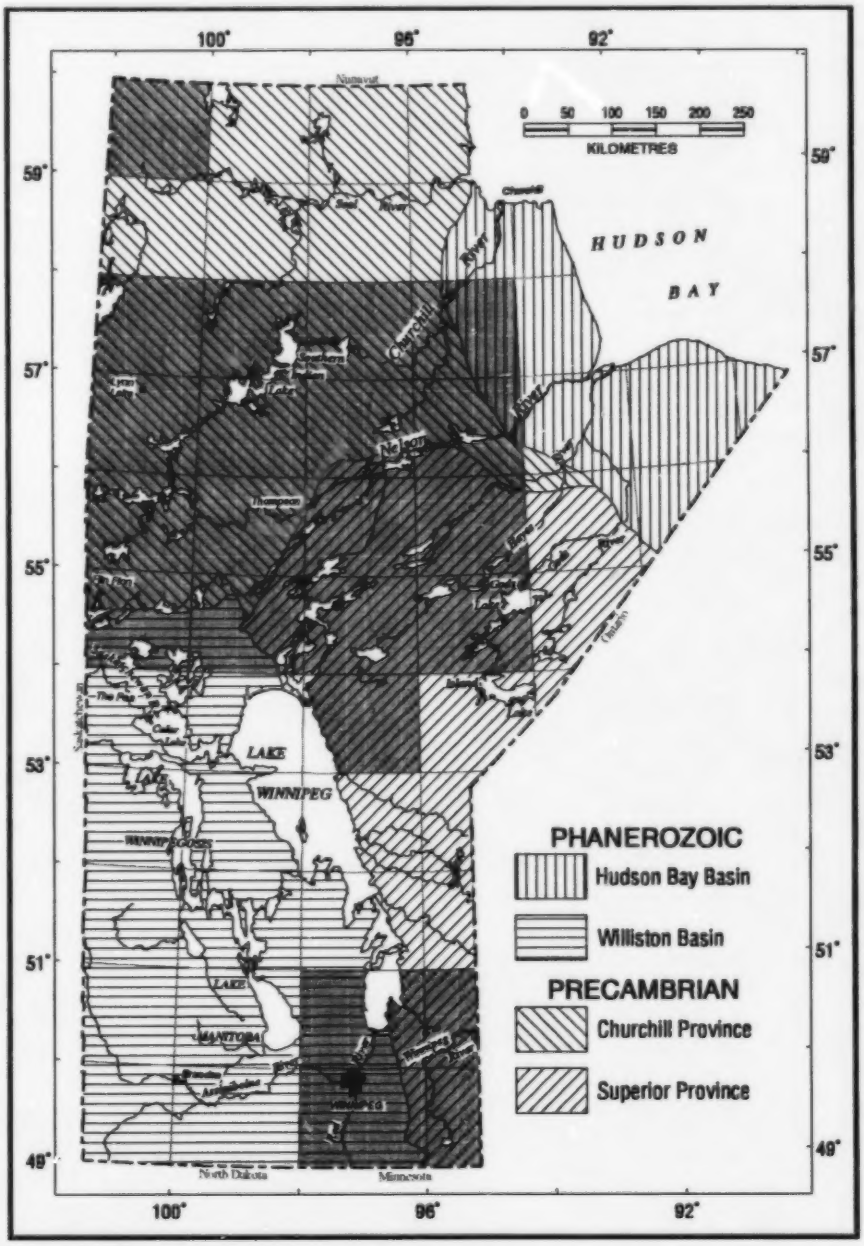

Figure GS-37-1: Areas of Manitoba covered by Bedrock Geology Compilation Map Series (BGCMS) maps.

INTRODUCTION

This year's work has focused on:

1) colour map production for the Cochrane–Seal rivers area (8 sheets), largely based on previously published 1:250 000 compilation maps (Schledewitz, 1986);
2) an update of BGCMS map 64B – Uhlman Lake sheet (Manitoba Energy and Mines, 1986), the original version of which is now out of print; and
3) development of a new BGCMS legend structure.

COCHRANE–SEAL RIVERS AREA

This area covers eight 1:250 000 scale NTS map sheets in northwestern Manitoba (Fig. GS-37-1): 54L (Churchill), 54M (Caribou River), 64-I (Shethanai Lake), 64J (Tadoule Lake), 64K (Whiskey Jack Lake), 64N (Kasmere Lake), 64O (Munroe Lake) and 64P (Nejanilini Lake). These sheets were first published in 1986 as part of Geological Report GR80-9 (Schledewitz, 1986). These eight BGCMS maps will represent digitization, refinement, conversion to BGCMS GIS format, editing and updating of the information from the GR80-9 maps and report. Newly released geophysical information (Viljoen et al., 1999) has been used for extrapolating dykes and contacts beneath lakes and overburden. Sheet 64N (Kasmere Lake) has been completed and released; the other seven sheets are in preparation, with 64O (Munroe Lake) and 64P (Nejanilini Lake) expected for release by March 31, 2001.

UHLMAN LAKE SHEET

The Uhlman Lake sheet (NTS 64B) has been updated to incorporate new information gathered during the last decade of field investigations, as well as the newly released geophysical information cited above. Release is expected by March 31, 2001. In the interim, in-house LaserJet™ prints of a scanned image of the out-of-print paper map (Manitoba Energy and Mines, 1986) are available for sale.

NEW LEGEND STRUCTURE

A new legend structure is now under development and will provide the following advantages over the existing structure:

1) There will be a more consistent legend across the entire province.
2) The new legend is not based on geological domains, which are often poorly defined.
3) There is no duplication of units cut by domain boundaries, resulting in a much simpler legend (the current legend has more than 1000 unit descriptions on 29 map sheets!).
4) Unit identifiers incorporate tectonostratigraphic as well as lithological information.
5) Unit identifiers are insensitive to text enhancements (e.g. bold face, subscript/superscript), polygon colour and patterning, and geological domain. This facilitates data exchange between various GIS/CADD platforms.
6) It will facilitate reclassification and other spatial analyses of lithological units using GIS (e.g. to produce an updated 1:1 000 000 bedrock geology map of Manitoba).

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GS-38 MINING SECTOR PARTICIPATION IN THE PROTECTED AREAS INITIATIVE

by P.D. Leskiw

Leskiw, P.D., 2000: Mining sector participation in the Protected Areas Initiative; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 225.

INTRODUCTION

The goal of World Wildlife Fund Canada's Endangered Spaces Campaign (ESC) is to establish a network of protected areas representing all of Canada's natural regions. Manitoba was the first province in Canada to commit to this goal. Manitoba's Protected Areas Initiative (PAI), in support of government commitments to the ESC, aims to protect representative samples of each of the province's terrestrial natural regions. Designation for protection to the ESC standards, and under Manitoba's PAI, prohibits all resource-extraction activity within the protected area, including mineral exploration and development. Protected areas may be designated as provincial parks or park reserves, wildlife management areas, ecological reserves, etc.

Candidate sites put forward by Parks Branch for consideration for protection have been designed using enduring-feature analysis. These Areas of Special Interest (ASI) are examined by the mining sector for mineral potential. The mining sector technical review is co-ordinated through the Mineral Exploration Liaison Committee (MELC) and its land-use subcommittee. Since 1997, this working group has reviewed sites and ASIs put forward by Parks Branch and made recommendations based on a ranking system designed by industry participants. The working group consists of representatives from mining and exploration associations in Manitoba, Parks Branch (Department of Conservation), Mines Branch and the Geological Survey (Department of Industry, Trade and Mines), and World Wildlife Fund Canada.

Industry participation and consultation for the review of Areas of Special Interest and sites for Manitoba's Network of Protected Areas provides Manitoba with increased certainty related to land-use issues.

PROGRESS

World Wildlife Fund Canada's Endangered Spaces Campaign ended in mid-2000. However, Manitoba's Protected Areas Initiative program has been extended for three more years. An additional 200 000 ha of the province were placed in protected status when Birch Island, Fisher Bay and Hudwin Lake Park Reserves were designated in the spring. Almost 40 000 ha of Wildlife Management Area lands were also placed in protected status. These actions bring to over 5.5 million ha the amount of land recognized under the Manitoba Protected Areas Initiative over the past decade.

The mining sector supported twenty-three sites for protection in May; in August, MELC ratified a further ten sites for protected status. Mineral rights within these new Rank One sites are under Withdrawal Pending status, while mineral rights within the designated protected lands are fully withdrawn. Rank One means the mining sector has confirmed the site for protection from exploration and mining.

The Manitoba Geological Survey provides technical support to the mining sector review of ASIs, by:

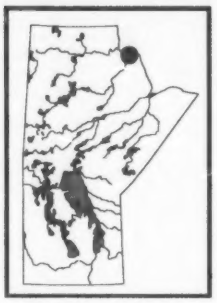
- providing regional and detailed geological and mineral-deposit information, as required, to gauge mineral potential in ASIs;
- performing mineral-potential assessments for all ASIs by summarizing relevant information from assessment reports, mineral-deposit and mineral-inventory reports, and other technical information;
- compiling and distributing digital geological and geophysical base maps on which mineral dispositions (e.g. claims, permits and leases) and mineral-occurrence information are represented; and
- representing the Department on the communications steering committee, whose purpose is to facilitate smooth operation of the mining sector review process.

In 2000, mineral resource assessments for 34 Areas of Special Interest, Wildlife Management Areas and proposed ecological reserves were generated by the Manitoba Geological Survey for review and potential protection.

GS-39

INDUSTRIAL MINERALS OF THE CHURCHILL AREA (PARTS OF NTS 54K/13SE AND 54L/16SW)

by J.D. Bamburak



Bamburak, J.D. 2000: Industrial minerals of the Churchill area (parts of NTS 54K/13SE and 54L/16SW); *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 226-230.

SUMMARY

The Churchill area has a recorded history of industrial mineral production since 1733. Current production is mainly for aggregates and minor building stone. Although current production is considerably less than for most southern Manitoba localities, it is still a significant area in the province because rail transportation costs would make alternatives prohibitively expensive. Unique fossils, such as trilobites, have been found along the shoreline of Hudson Bay. Future exploitation may involve quartzite production as a source of silica, production of rare elements from quartz veins, and minor rock collecting.

INTRODUCTION

Industrial Mineral Commodities

Past and present production of industrial mineral commodities in the Churchill area has been mainly for aggregate purposes, such as road construction and maintenance. Sand and gravel, as well as crushed bedrock, were and are in production at numerous locations. The sand and gravel production is not covered in this paper. Production of building stone was carried out more than 250 years ago. Current production is limited, but is being carried out on one project.

Aggregate and, to a lesser extent, building stone are valuable commodities in the Churchill area, considering the long distance to equivalent materials, which can only be provided by rail.

Previous Investigations

In 1879, the Churchill area was visited by Bell (1881) of the Geological Survey of Canada (GSC), who documented the bedrock and surficial geology for the first time. He recognized an anticlinal structure in the vicinity of Churchill Harbour, and recommended that the term 'Churchill Quartzite' be used (Bostock, 1969). Tyrrell (1898) of the GSC described the glacial geology of the Churchill area and made some bedrock observations in 1893. He observed ripple marks and abundant crossbeds in the Churchill Quartzite (Bostock, 1969).

Williams (1948) spent a month at Churchill in 1947. He noted that the Churchill Quartzite commonly consists of 70% well rounded quartz and 30% interstitial sericite. Williams reinterpreted the structure at the mouth of the Churchill River to be a syncline and suggested that Silurian rocks underlie the Churchill area (Bostock, 1969). In 1967, Bostock mapped the Churchill area as part of Operation Winisk, a GSC helicopter-supported mapping project covering the Hudson Bay Lowlands. Table GS-39-1 is derived from his report (Bostock, 1969).

Table GS-39-1: Table of formations, Churchill area.

Formation/Member	Maximum thickness (m)	Lithology
Silurian:		
Severn River	150	Dolomitic limestone, with up to 25% quartz sand in lower beds
Ordovician:		
Churchill River Group	90	Dolomite and limestone
Precambrian:		
Churchill Quartzite	730	Quartzite, subgreywacke

In 1974, Schledewitz of Manitoba Energy and Mines mapped the extent of the Precambrian bedrock in the Churchill area. The Precambrian and Phanerozoic aggregate resources were documented by Young (1982a, b). Watson (1985) described sampling and analyses undertaken on the Churchill Quartzite.

The Manitoba Museum of Man and Nature and the Royal Ontario Museum have collected fossil specimens along the tidal flats east of Churchill for a number of years (Young et al., 1999).

Current Investigations

Precambrian and Phanerozoic bedrock outcrops and quarry locations in the Churchill area were inventoried over a nine-day period in July (Fig. GS-39-1). Introduction to the geology of the area and efficient access to outcrops and quarries were provided by staff of the Manitoba Museum of Man and Nature and the Royal Ontario Museum.

Dorne Lindal of the Manitoba Geological Survey is presently compiling a bedrock geology compilation map for NTS 54L, which includes part of the Churchill area (Manitoba Industry, Trade and Mines, in prep.).

GEOLOGY

The Precambrian and Paleozoic geology of the Churchill area is shown in Table GS-39-1. The Precambrian Churchill Quartzite is unconformably overlain by the Ordovician Churchill River Group, which is overlain by the Silurian Severn River Formation.

Precambrian Churchill Quartzite forms a low ridge rising above dipping Ordovician Churchill Group beds in Figure GS-39-2. The Ordovician beds were draped over the flanks of the quartzite, which had been eroded prior to deposition of the younger sediment. The breccia clasts spalled from the quartzite were cemented by the carbonate infill (Fig. GS-39-3). Most of the Ordovician and Silurian beds have been removed from the tops of the Churchill Quartzite ridges. However, in places it is still possible to see remnants of the younger beds preserved in deep fissures (Fig. GS-39-4).

Pleistocene glacial sediments and Recent alluvium and bogs blanket the Churchill area, except near the shoreline of Hudson Bay.

CRUSHED STONE

Precambrian Churchill Quartzite

Five quartzite samples were collected as a potential source of aggregate in the Churchill area by Young (1982a, b). The samples were from the Precambrian Churchill Quartzite that outcrops for 25 km along the shoreline of Hudson Bay and in the mouth of the Churchill River. Four of the samples met the specifications for base course, traffic bituminous, concrete, ballast and terrazzo aggregate.

Recently, Churchill Quartzite was quarried on the west side of the Churchill River, south of Fort Prince of Wales, to construct a weir. The purpose of the weir was to pond fresh water for drinking purposes. A small quartzite quarry has been opened near the shoreline northeast of Churchill by Mervin Walkoski, a local resident.

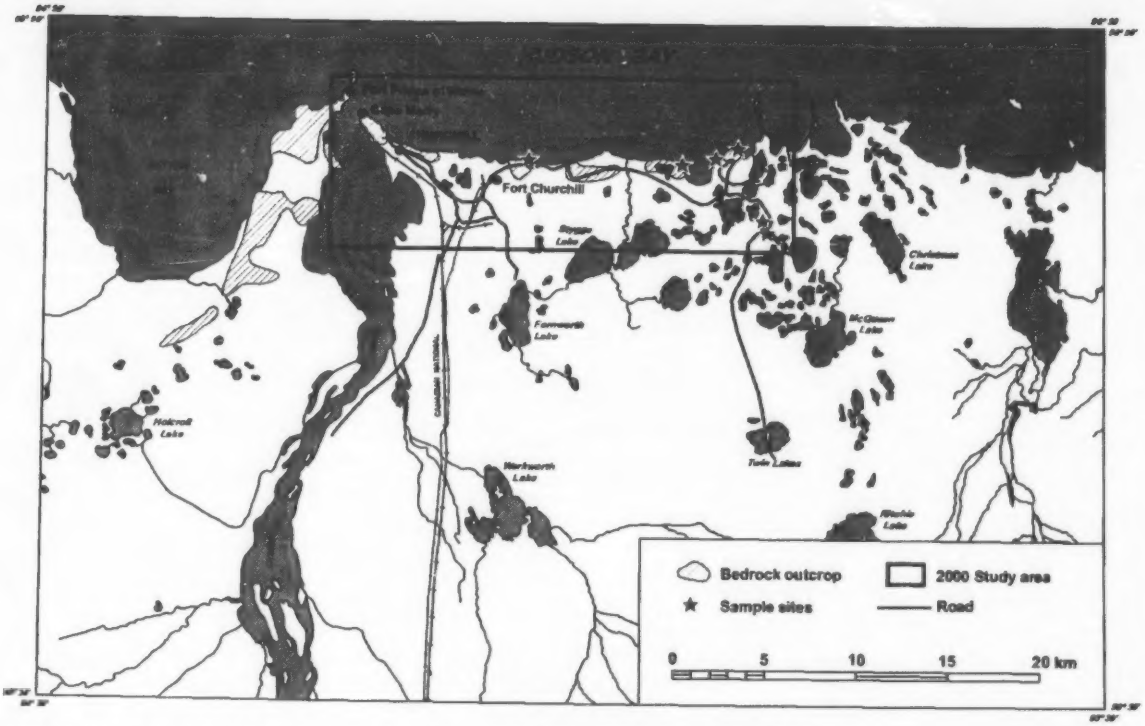


Figure GS-39-1: Location of the Churchill area in northern Manitoba, with outcrop sampling sites.

Figure GS-39-2: View looking east along Hudson Bay shoreline, east of Churchill. Dark-coloured Precambrian Churchill Quartzite (in backgound) forms a low ridge above light-coloured, dipping Ordovician Churchill Group dolomite (in foreground). Note individuals near 'bergy-bit' in upper right of photo for scale.

Ordovician and Silurian Dolomite and Limestone

Ordovician and Silurian carbonate rocks were not evaluated as a potential source of aggregate by Young (1982a, b). Mervin Walkoski is currently producing aggregate from quarries within Ordovician and Silurian beds in the vicinity of the above quartzite quarry.

BUILDING STONE

Fort Prince of Wales and Cape Merry

According to Tyrrell (1898), the Hudson's Bay Company began construction of Fort Prince of Wales in 1733. The fort was built on the summit of the low point to the west of the entrance to Churchill Harbour (Fig. GS-39-1). Churchill Quartzite was quarried a short distance to the west of the fort, along the shore. To prevent the cannons on the top of the fort walls from recoiling into the courtyard, the walls of the fort were made 12 m thick (Fig. GS-39-5). Some field stone was also used, in the construction of the interior walls. In the past few years, stone has been quarried to rebuild portions of the outer walls of the fort.

The Hudson's Bay Company also built a cannon wall in the 1700s at Cape Merry, using a similar construction method to that used at the fort. The Cape Merry battery was constructed on the east side of the entrance to Churchill Harbour to ensure that enemy vessels could be caught in a crossfire.

Fort Prince of Wales and Cape Merry are both preserved as Historic Sites by the Federal Government. These sites are popular tourist attractions (Fig. GS-39-5).

Figure GS-39-3: Dark-coloured quartzite breccia clasts within light-coloured Ordovician dolomite.

Figure GS-39-4: Silurian Severn River remnant, with quartzite clasts, within a light-coloured, smooth-weathered, Precambrian Churchill Quartzite fissure.

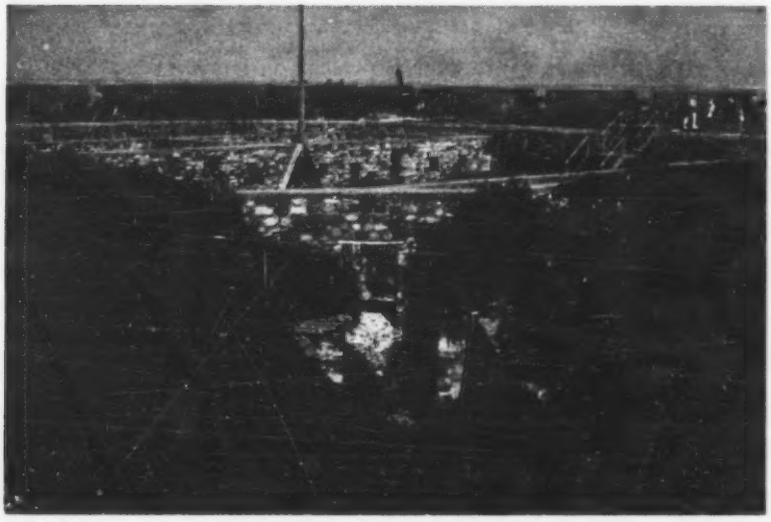


Figure GS-39-5: Fort Prince of Wales, interior view looking northwest across courtyard. Note tourists in upper right of photo for scale.

Brian Ladoon's Hotel

Brian Ladoon, a local resident, is constructing a hotel from field stone and Churchill Quartzite (Fig. GS-39-6). According to the Winnipeg Free Press (Jan. 12, 1997), the hotel was scheduled for completion by August 1999 at a cost of $700 000–900 000.

SILICA SOURCE

Watson (1985) analyzed eleven Churchill Quartzite samples, collected by Schledewitz in 1974. He concluded that most of the samples, which ranged from 75.50 to 97.65% SiO_2, met the specifications for silicon carbide, ferro-silicon and various fluxes. The main impurities in the quartzite are mica and minor feldspar.

MINERAL AND FOSSIL COLLECTING

Lazulite

The first recorded occurrence of lazulite (hydrated Mg-Fe-Al phosphate) in Canada was made in 1879 by Bell (1881) at a locality situated about 3 km east of Churchill (Hoffman, 1881; Bostock, 1969). Pegmatitic lazulite and chlorite are present as rare accessory minerals in quartz veins that intruded the Churchill Quartzite, according to Schledewitz (1986). The lazulite is blue, massive and forms irregular masses up to 4 cm across. The most abundant lazulite can be found on the east bank of the Churchill River in the vicinity of Cape Merry.

Specular Hematite

Bostock (1969) noted that quartz veins in the Churchill Quartzite (his 'subgreywacke') had local concentrations of specular hematite. Laminae of specular hematite, up to 1 cm thick, were also observed to locally form 4 to 5% of the subgreywacke. Schledewitz (1986) also reported coarse-grained specular hematite within 1 to 2 mm wide quartz veins that intruded the Churchill Quartzite. Like Bostock, he found specular hematite along the bedding planes and as disseminated grains in the quartzite.

Samples of specular hematite were collected in July 2000 for trace-element analysis.

Pyrite

Bell (1881) reported tiny pyrite grains in some veins and green Cu carbonate in one. Bostock (1969) observed finely disseminated pyrite, apparently associated with fractures in the Churchill Quartzite, near the east margin of exposure in the Churchill area.

Well developed cubes of pyrite can be found within the Ordovician Churchill Group. The largest cubes are about 1 cm on a side and are usually found in a larger mass of sulphide. Pyrite clots are also present in the overlying Ordovician Severn River Formation, where the beds are draped directly over the Precambrian Churchill Quartzite. Samples collected during July 2000 will be tested to determine if sulphide minerals, other than pyrite, are present.

Ordovician and Silurian Fossils

A 70 cm long by 30 cm wide trilobite was found on the tidal flat near Churchill in July 1998 by Dave Rudkin of the Royal Ontario Museum. It is the largest recorded complete fossil of a trilobite ever found (Winnipeg Free Press, May 27, 2000). The trilobite was within the Ordovician Churchill River Group. In 1985, a smaller 42 cm long trilobite was also found. A small trilobite found during July 2000 is shown in Figure GS-39-7. Other fossils preserved at the site include conodonts, aulacerid stromatoporoids and branching tabulate coral (Young et al., 1999).

In the same general area, Silurian fossils have been located in the Severn River Formation. These include stromatoporoids, corals, cephalopods and brachiopods (Young et al., 1999).

ACKNOWLEDGMENTS

The assistance of the staff of the Manitoba Museum of Man and Nature and of the Royal Ontario Museum is gratefully acknowledged. The locations they provided for the bedrock exposures on the tidal flats saved countless hours in determining outcrop availability and access. In particular, I would like to acknowledge Dave Wright and Ed Dobrzanski for acting as field assistants and watching for polar bears.

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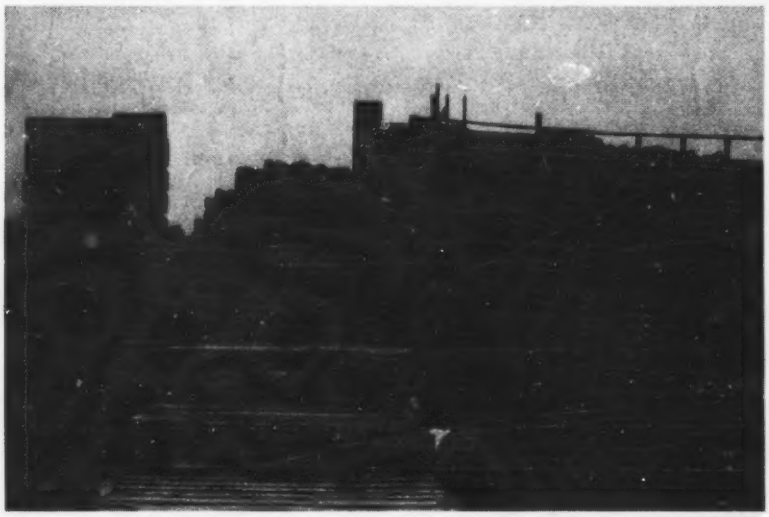

Figure GS-39-6: Brian Ladoon's hotel site, front entrance.

Figure GS-39-7: Trilobite fossil, found July 2000.

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GIS-BASED INTERACTIVE MAPS ON THE WEB

by M. Pacey

Pacey, M. 2000: GIS-based interactive maps on the Web; *in* Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 231.

SUMMARY

Web-enabled Geographic Information System (GIS) technology is a tool used to present geospatial data on the Internet. This technology provides the Internet client with 24-hour access to the Manitoba Geological Survey's information. To access the database, an Internet connection and a Java-enabled Web browser, such as Microsoft Internet Explorer or Netscape Navigator, are required. The website has a user-friendly GIS style, look and feel, with tools that allow query and analysis of the most current datasets relevant to the mining and mineral-exploration community. The 'GIS Map Gallery' is the access point for information on mineral claims, assessment files, orthophotography and geological databases. The GIS Map Gallery can be found at:

http://www.gov.mb.ca/em/geoscience/gis/gis-index.html

INTRODUCTION

Our clients have asked for a service that allows quick retrieval of the information essential to their business. The mining and mineral-exploration community operates worldwide, from the smallest northern settlement to the largest urban corporate headquarters. In response to our clients' needs, we have created the GIS Map Gallery, an easy-to-use, interactive, information reference mapping tool. This tool is provided via the Internet to accommodate both our most isolated clients and our centralized ones. The Internet has provided the medium for instant access to information from almost anywhere in the world. Prospectors, geologists, mining company staff and other interested stakeholders are now able to:

- see where mineral dispositions are located and what infrastructure exists in their area of interest;
- determine areas that are restricted in terms of staking and exploration;
- find out who holds what ground;
- find out what ground is available for staking in their area of interest;
- find out where recent staking has taken place; and
- find out what work was done in the area.

The graphic interface will aid clients in preparing exploration programs and in the decision-making process.

A NEW MINING AND EXPLORATION TOOL

The Web-enabled GIS Map Gallery is an interactive mapping system (Fig. GS-40-1) that allows users to zoom in and out, pan around the map, download reports, execute queries on databases, hyperlink to other websites, and print maps. Included is a detailed help facility that guides the visitor through the use of the navigation and query tools found in the interface. The databases are kept current and relevant to the activities of our clients. The system includes the following databases:

- Mining claims
- Assessment files
- Mineral leases
- Land-use information
- Orthophotography
- Exploration permits
- Quarry leases
- Patent claims
- Road network
- Bedrock geology

In addition to querying and selecting items from these data sets, the user is able to execute true GIS-type spatial queries, such as:

- Display mining claims that are within a certain distance of road access (by creating a buffer zone around the chosen road feature).
- Display all assessment files available in a selected NTS area.
- Highlight all mining claims due to expire within the next ten days, using the 'Search' tool.

Grid and UTM co-ordinate reference, distance measurements and sophisticated SQL (Structured Query Language) queries are also available within the interface. The completed analysis can be captured by printing out a reference or presentation copy of your work from your Internet browser.

Planned enhancements to the GIS Map Gallery include published documents from geological mapping projects, the ability to hyperlink from the map to download assessment files, and the ability to obtain publication information for the province. The ultimate goal is to enable the client to determine exploration targets in the privacy of home or office, with the help of the Manitoba Geological Survey's GIS Map Gallery.

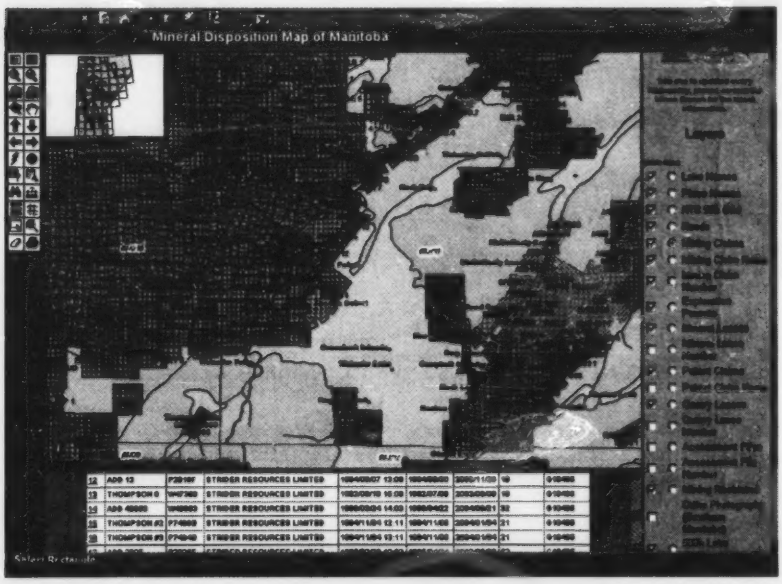

Figure GS-40-1: Screen capture from the GIS Map Gallery interface.

ACKNOWLEDGMENTS

Many thanks to Mike Fedak and Yong Zhang for their tireless updates of the mineral claims database.

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2000S-1 Geology of the Kistigan Lake area (NTS 53K/10)
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2000S-4 Aswapiswanan Lake east area (part of NTS 53L/6NW)
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